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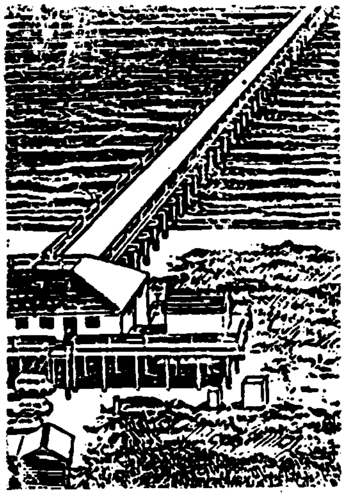

COASTAL GEOLOGY AND GEOTECHNICAL PROGRAM

2

TECHNICAL REPORT CERC-92-4

GEOMORPHIC VARIABILITY IN THE COASTAL ZONE

by

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The coastlines of the world's oceans encompass a tremendous variety of geomorphic and geologic structures. They range from rocky cliffs to sandy barrier beaches to low-lying swampy wetlands. The geomorphic forms were created by the interaction of antecedent geology, physical dynamic processes, and man-made intervention. Variable features are usually composed of unconsolidated materials that respond rapidly to changes in the dynamic environment. More stable features are usually associated with consolidated rock or occur in quiescent environments. The geologic history of shorelines can be inferred from a careful study of geomorphic structures, coupled with additional data on physical processes and historic events. Many of the study techniques are relatively simple, consisting of analysis of existing maps and historical sources.

An understanding of the processes which have shaped the shore is crucial to the design of coastal structures and to the intelligent management of coastal resources and habitats. In addition, understanding of the form/process relationships between geomorphology and dynamics may allow coastal scientists to more accurately predict the results of construction or other modifications along the shore.

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Preface

The study reported herein results from research performed by the Coastal Engineering Research Center (CERC) of the US Army Engineer Waterways Experiment Station (WES) under Work Unit No. 32538, "Survey of Technologies in Coastal Geology," Coastal Geology and Geotechnical Program, authorized by the US Army Corps of Engineers (USACE). Mr. John Sanda was USACE Technical Monitor. Ms. Carolyn M. Holmes was the CERC Coastal Program Manager.

This report was prepared by Dr. Joann Mossa, Department of Geography, University of Florida, under the US Army Summer Faculty Research and Engineering Program and by Messrs. Edward P. Meisburger and Andrew Morang, Coastal Geology Unit, Coastal Structures and Evaluation Branch (CSEB), Engineering Development Division (EDD), under the general direction of Mr. Thomas W. Richardson, Chief, EDD, and Ms. Joan Pope, Chief, CSEB. Director of CERC during the investigation was Dr. James Houston, and Assistant Director was Mr. Charles Calhoun, Jr. Peer review was provided by Dr. Donald Stauble, Ms. Joan Pope, and Mr. J. B. Smith, WES.

At the time of publication of this report, Director of WES was Dr. Robert W. Whalin. Commander and Deputy Director was COL Leonard G. Hassell, EN.

Conversion Factors, Non-SI to SI (Metric) Units of Measurement

Non-SI units of measurement used in this report can be converted to SI (metric) units as follows:

Multiply	By	To Obtain
millibars	10^{-1}	kilopascals
miles per hour	1.609347	kilometers per hour

1 Introduction

Background

Many geomorphic features are relatively stable and changes in their form occur slowly; others are less stable and are subject to comparatively rapid change in response to dynamic environmental factors. Variable features are usually composed of unconsolidated or friable materials that react rapidly to changes in the dynamic environment. More stable features are usually associated with consolidated rock and/or the absence of potent geological processes.

Few geological environments contain a greater variety of dynamic geomorphic features than the coastal zones of oceans and large lakes. Because of the many variable features, complex processes, and submerged areas, study and description of coastal areas present special problems that do not arise when examining more stable environments. In general, study of coastal areas requires a process-oriented approach and a larger database than are usually needed for study of more stable and accessible environments. The large database is needed for two principal reasons. First, time series data collection on many variable geomorphic features is needed to determine the range of temporal variability. Second, selection of representative data points and projection of data by visual observation and aerial photography are not possible in submerged parts of the coastal zone. Consequently, a dense data matrix is needed for an adequate and reliable representation of bottom morphology and sediment distribution.

Temporal variations in coastal geomorphology occur in: (a) cyclic patterns; (b) as a result of intermittent noncyclic events; or, (c) as long-term trends. Cyclic variations occur on a periodic or repetitious basis and are generally related to processes like astronomical tide and seasonal sea and swell patterns. Intermittent events are noncyclic occurrences such as large storms or earthquakes. They are of relatively short duration but often have lasting consequences for coastal morphology. Long-term trends include gradual changes in relative sea level or climatic patterns which, in turn, cause slow but often significant changes in coastal features or processes. Over the time scales of modern process studies, they are considered noncyclic, although over geologic time scales they may have a periodic component. The detection of such long-term trends is often difficult, but they may be discovered by

comparative analysis of historical maps, charts, and aerial photographs that show changes over decades or centuries.

Scope

This report identifies and briefly discusses variable coastal features, the processes and factors that affect them, and the available technology and techniques for their study. The features discussed here are, for the most part, associated with the direct effects of marine and lacustrine processes. Terrestrial features and processes that occur inland of the coastline are discussed briefly in the following section. They are described in greater detail in many standard texts dealing with geomorphology, physical geography and physical geology (Chorley, Schumm, and Sugden 1984; Strahler and Strahler 1987; Bloom 1991).

Coastal Classification

Shorelines are influenced by a combination of nonmarine and marine processes. Shepard's (1963) genetic classification of coasts reflects the dichotomy of nonmarine and marine influences. Primary coasts, according to his classification, are essentially nonmarine in origin and are shaped mainly by land erosion, subaerial deposition, volcanism, and diastrophism. Secondary coasts are shaped by marine processes, including wave erosion, marine deposition, and organic/biologic modification.

Primary shorelines have yet to be appreciably modified by marine processes. Land erosion coasts include: (a) drowned river valleys; (b) drowned glacially eroded coasts, including fjords; and (c) drowned karst topography. Coasts of subaerial deposition include: (a) those principally reflecting river deposition with either deltas of various forms or alluvial plains; (b) those reflecting glacial deposition, with partially submerged moraines, drumlins, or drift features; (c) wind deposition with active or fossil dunes, or sand flats; and (d) those reflecting landslide deposition and perhaps other forms of mass wasting. Volcanic coasts include those influenced by lava flows, tephra or volcanic ash, and volcanic collapse or explosion. Coasts may also be shaped by diastrophic movements, which produce faulted and folded coasts of various types, and coasts with sedimentary extrusions such as salt domes and diapirs.

Information Sources

The most direct and accurate means of obtaining information on the variability of coastal features in an area is by conducting repetitive surveys over a

period of time. This period must be sufficient to cover the main cyclic events associated with the seasons. Repeated surveys should cover each of the seasons of a yearly cycle. In addition, survey density and timing suitable to document the effects of extreme events, such as major storms, floods, and hurricanes, are also desirable. Study of long-term cycles by repetitive surveys is not widely practiced because of the expense and continuing commitment needed to obtain sufficient data.

Historical maps, charts, and aerial photographs are important sources of geologic information. A valuable source of these items is the National Ocean Service (NOS) and its predecessors. Survey data extending back 150 years are available for many areas, and preliminary charts of considerably larger scale and detail than the published versions can be obtained from the NOS archives. These historic sources are of great value in revealing long-term trends. Some sources for historical and recent aerial photography and maps are listed in Fulton (1981).

Data on intermittent noncyclic events are usually difficult or impossible to obtain. In particular, baseline data of conditions immediately before the event are often nonexistent because there is typically insufficient advance warning to organize and conduct an adequate survey. However, aerial photographs can often be obtained even with fairly short notice and are valuable in showing conditions before an event, especially in the shore zone. Post-event surveys should be conducted as soon as the affected area becomes accessible.

Information regarding past events can often be found through a combination of field and laboratory tests. These include a variety of stratigraphic and sedimentologic techniques, such as the analyses of sedimentary structures, grain size, sediment composition, heavy minerals and fossils, and the radiometric dating of organic remains in sediment deposits and of other isotopes of natural and human-derived sources. These techniques, in addition to archeologic and pedologic ones, are of value in working out past environmental changes and events over a variety of time scales, although some parts of the geologic record may have been obliterated.

Literature sources on coastal features and processes, many containing historical information, are abundant, but a considerable research effort is needed to cull out pertinent works. In a recent computer-assisted literature search, over 1,400 items were listed under the key words "coastal geology," "coastal geography," "coastal geomorphology," and "coastal classification." Most had been published since 1970. Selected key sources of coastal engineering information, including meteorology and climatology, water levels, wave and currents, ice, beach erosion and littoral transport, topography and bathymetry, and geology and geomorphology, are described in Chu, Lund, and Camfield (1987). Historical geomorphic data can also be acquired from a variety of sources (Trimble and Cook 1991). Local records such as tax assessments, deeds, and local surveys may also be useful (Fulton 1981). Furthermore, there are many unpublished sources of coastal information that can be sought from government agencies, universities, and private firms.

2 Relevant Processes and Factors

From both a temporal and spatial perspective, the geomorphic variability of a coast reflects a balance between forces that promote change and materials that resist change. Forces promoting change include atmospheric, oceanographic, biologic, and terrestrial processes that act individually and in combination. Some of these are important locally and others are important regionally, over short- and long-term periods. The processes and products of the interaction of the various factors in coastal environments are complex, and it is often advantageous for a researcher to adopt a holistic approach.

Because the coast lies at the boundary between the atmosphere, ocean, and land, processes within each of these environments are important in promoting coastal change. In the atmosphere, factors operating on varying spatial and temporal scales include climate, wind, and cyclonic disturbances. Oceanic processes that influence coasts include waves, tides, currents, storm surges, tsunamis, and sea level (eustatic) changes. Terrestrial processes may be internally driven, such as by tectonic factors, or externally driven, by climatic and meteorologic agents such as running water, groundwater, and ice. Biologic processes may also come into play, influencing both the shape and material of coastal environments. Interactions between many of the major processes are described briefly in subsequent pages.

Materials in coastal environments have varying properties, which characterize their ability to resist deformation by stresses, their ability to resist weathering and abrasion, and their ability to resist transportation by a fluid agent. The strengths of materials depend on the nature of the solids that comprise them, the nature of the voids (whether filled, or partly filled, with water or with air), and the forces holding the aggregates together. Some factors that characterize resistance to shear include rock strength, state of weathering, and the spacing, width, and continuity of joints and voids. Frictional and cohesive strength, mineral hardness, and fabric also cause materials to vary in their resistance to erosion by abrasion, fluid stresses, corrosion, and plucking. Materials vary in their resistance to entrainment and transportation by fluids according to their volume, density, and friction with the bed. Because coastal sediments display a range of characteristics from consolidated to unconsolidated, gravel to clay, terrigenous to biogenic, they show great variety in their resistance to natural forces. The nature and

resistance of inland materials, which can be transported to the coast by terrestrial processes, also contribute to the geomorphic variability of coasts.

Climate

Climate and weather are terms used to describe a broad group of interrelated atmospheric processes. Weather characterizes the overall state of the atmosphere on short time scales of minutes to months. Climate, in contrast, characterizes the long-term conditions of an area, using averages together with measures of dispersion and frequency.

Several interrelated climatic factors such as temperature, pressure, wind, and moisture have a great influence on the development of the coastal zone. These factors may change as long-term trends, and also display cyclic and noncyclic variations. Distinctive cyclic variations include seasonal and daily changes associated with the revolution and rotation of the earth. Regional variations in paleoclimatic and modern climatic factors also contribute to the spatial variability of coastal geomorphic features. Of the various climatic factors, wind most strongly influences modern process and geomorphic variability of coasts, both directly and indirectly. Wind is discussed subsequently in a separate section.

Evidence from stratigraphic, paleontologic, palynologic, radiometric, pedologic, and archeologic data shows that many climatic changes of global and regional scale have occurred throughout geologic time. While the long-term paleoclimatic history of the earth is not well known, paleoclimatic data assist in characterizing general aspects of the many profound and widespread changes that have occurred throughout geologic time.

Most evidence indicates that over the last 500 million years, the earth was warmer than present, with the poles being ice-free most of this time (Frakes 1979; Lamb 1982). About 55 million years ago, global climate began a long cooling trend, which led to the development of the high-latitude ice sheets.

Superimposed upon this trend, over the last 1 to 2 million years during the Quaternary period (Pleistocene and Holocene epochs), are a number of alternating glacial and interglacial episodes. In the Northern Hemisphere, these episodes were characterized by widespread waxing and waning of the ice sheets into the mid-latitudes. Falling and rising temperatures in both hemispheres accompanied glacial advance and retreat. During cooler periods, more widespread glaciation of high altitudes was also common.

These climatic changes were accompanied by large fluctuations in sea level. During glaciations, water was withdrawn from the ocean basins and stored in ice sheets and mountain glaciers. During interglacial periods, it returned to the ocean basins. Many of the large inland lakes also may have

expanded during the colder climates of glaciations, perhaps in response to reduced evaporation. Crustal depression occurred with ice advance over the continents; crustal rebound followed deglaciation, further complicating the history of relative sea level changes along coastlines.

Because of changes in relative sea level, even those coasts not directly glaciated have been affected to some degree. But coasts were affected in differing ways by these processes and events, adding to their geomorphic variability. Such eustatic, climatic, and crustal changes are still occurring to varying degrees in different parts of the world. This subject will be further discussed in the section on relative sea level later in this part of the report.

Variations in paleoclimate and glacial activity account for much of the variability in geomorphic landforms and processes along coasts. Examples of erosional coasts include fjords, which were produced by the erosive activity of mountain glaciers in conjunction with relative sea level changes. The erosive activity of the mid-latitude North American ice sheets was also integral in the formation of the Great Lakes basins (Hough 1968). As characterized by Shepard (1963), depositional glacial coastlines may consist of tills, moraines, and drumlins. The north shore of Long Island Sound, extending from Connecticut to Massachusetts, displays these features.

Modern climatic conditions have been well-documented by systematic and accurate meteorological observations in many parts of the world. In some places, such records exist for periods as long as 2 centuries. These form a foundation for a reasonably accurate model of the earth's present climatic patterns and of anticipated trends. The use of satellites for remote sensing of the atmosphere and the development of general circulation models are assisting in the interpretation of past and the prediction of future climates.

Coasts are directly and indirectly affected by modern climatic conditions. Differences in temperature and precipitation influence organisms, vegetation type, and biomass production (Whittaker and Likens 1975). Climatic variation also affects the type and intensity of weathering and the rates of denudation (Langbein and Schumm 1958; Knighton 1984), which in turn influence geomorphic features along coasts. Atmospheric conditions are intimately linked to oceanographic conditions, so that a change in one will cause a change in the other.

Climatic factors need not operate directly in the coastal zone to be relevant. For example, wind-generated waves in remote offshore areas can eventually reach and affect the coast, and precipitation in inland areas is important in producing and transporting sediment that enters the coastal system. The assessment of climatic factors therefore requires consideration of climatic data from outside, as well as within, the coastal zone.

Wind

Wind has a considerable influence on the coastal zone. This influence may be direct, since wind is an agent causing erosion and transportation. Near the coast, winds often have a predominant onshore-offshore orientation. If winds are blowing offshore, sediments may be eroded and transported from inland areas and then deposited in the coastal zone. Sediments from within the coastal zone, particularly the dry sand of dunes and beaches, may be eroded and transported offshore directly by wind.

The influence of wind is also indirect because wind stress causes the formation of waves and oceanic circulation. Over large areas and long time scales, predominant wind velocities and durations affect wave climate, whereas over shorter time scales, storm winds generate waves capable of great geomorphic activity. Waves generated by wind, in turn, often approach the coast at an angle, leading to longshore currents. Strong onshore winds, if sustained, may also cause increased water levels or setup.

Wind is caused by a pressure gradient, or horizontal differences in pressure across an area. This, in turn, may be created by differences in temperature, as the pressure, temperature, and density of gases are interrelated. Depending on the scale of the pressure and temperature variations, atmospheric phenomena range from those of global scale, which are generally persistent, to those of local scale and short duration, such as storms.

On a global scale, the wind systems occur in characteristic belts, being named for the direction from which they blow. Near the earth's surface, the zone about 5° to 25° poleward of the equator is characterized by the northeast trades in the Northern Hemisphere, and by the southeast trades in the Southern Hemisphere. Wind direction is predominantly westerly between about 30° and 60° in both hemispheres, whereas easterly winds are predominant poleward of 60°. While these belts shift somewhat over the year, these create characteristic wind patterns and intensities, which affect wave energy and predominant wave directions, oceanic circulation, and sediment transport by both wind and water.

Seasonal shifts in wind also occur in some regions, notably in southeast Asia where monsoon winds occur. The monsoon brings strong seasonal onshore winds, which influence the wave environment and circulation of the Indian Ocean (Davies 1980). The monsoon is accompanied by abundant precipitation caused by moist maritime air masses. Other local winds of note include the katabatic winds in cold climates, mountain and valley winds, and the sea and land breezes (Hsu 1988).

The sea breeze and land breeze are common phenomena that occur because of diurnal temperature differences between land and water. During the day, especially in the summer, the land warms more rapidly than the water. The

air expands and rises as it warms, forming a belt of lower pressure along the coast. The pressure gradient between the water and land causes a sea breeze in which air from the sea blows landward. The opposite happens at night. The land cools more rapidly because it is less efficient than water at storing heat. The air over the sea has lower pressure, resulting in vertical convection. The convection causes the land breeze to flow seaward.

Cyclonic Disturbances

Centers of relatively low atmospheric pressure, also known as cyclones, are associated with windy, cloudy, and wet weather. In contrast, anticyclones, or high pressure weather patterns, are generally associated with calm, dry weather. Cyclones occur in many sizes and forms, including continental-size extratropical cyclones, tropical cyclones of varying intensity, and tornadoes. All are capable of causing significant geomorphic change, because of the winds, waves, and storm surges associated with them. The intensity of cyclonic winds is controlled by the pressure gradient, or change in pressure measured along a line at a right angle to the isobars. Low pressure may also occur as elongated troughs, which range in scale from a few hundred kilometers to continental.

Extratropical cyclones, also known as wave or mid-latitude cyclones, are transient features that develop in various stages. Initially, they develop along the polar front, a narrow zone separating the cold, polar easterly winds from the warmer, mid-latitude westerly winds. A wave or indentation forms between the cold and warm air masses, causing the cold air to invade warmer territory and the warm air to enter colder territory along sharply defined fronts. The cold air masses, being denser, force up the warm air as they move along the surface, whereas the lighter warm air masses move over the colder air masses. The cold air masses eventually dominate the surface after all of the warm air is forced off the ground. This cuts off the source of moisture and energy, causing the system to die.

Central pressures in extratropical cyclones vary greatly, with lows of 940 mb* (compared to average sea level pressure of 1013.2 mb), although typically they do not fall below 980 mb. In both hemispheres, the most intense extratropical cyclones occur in the winter, with less intense systems developing in fall and spring. Cyclonic winds blow counterclockwise in the Northern Hemisphere, and clockwise in the Southern Hemisphere. Thus, depending upon the relationship of the storm orientation and track to the coast, there are several possible wind sequences (e.g., onshore winds followed by offshore winds) that affect wave, current, and sediment transport patterns (Niedoroda et al. 1984). Typically, the systems are transported with the westerly flow and move toward the east, although their tracks depend on

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* **A table of factors for converting non-SI units of measurement to SI (metric) units is presented on page x.**

upper air circulation. They usually pass an area within a day or two, although sometimes they remain stationary over an area for several days.

Tropical cyclones, in contrast, are believed to originate from preexisting disturbances called easterly waves, which are found in the tropical oceans in the 8° to 15° latitude band. The conditions that promote their formation include low level cyclonic vorticity, a Coriolis force, minimal vertical changes in velocity, sea surface temperatures warmer than 27°C, unstable lapse rates, and high humidities at mid-tropospheric heights (Riehl 1979). Such disturbances do not form in oceans such as the southern Atlantic, which do not meet these conditions.

If the storm intensifies and average wind speed increases, tropical cyclonic disturbances become tropical depressions, tropical storms, and finally hurricanes, with average wind speeds in excess of 75 mph. Increasing wind speeds are caused by an increased pressure gradient, where central pressures in severe hurricanes can fall below 900 mb. The great majority of tropical cyclones develop in the late summer and early fall in both hemispheres. As with other cyclones, meteorologic events follow a characteristic sequence (Figure 1) that affects coastal geomorphic features, depending upon the relationship of the storm orientation and track to the coast (i.e., Penland and Suter 1984).

The geomorphic activity of tornadoes and waterspouts, because they are small in size and develop and die rather quickly, is not on the same scale as other cyclones. Tornadoes are associated with cumulonimbus or thunderstorm clouds, and are most common in zones where air masses of contrasting temperature and moisture meet, such as the interior of the United States. Occasionally, tornadoes and waterspouts affect coastlines, with most damage being caused by high wind speeds; wind duration and fetch are typically insufficient to develop large waves.

The relative influence of extratropical cyclones compared to tropical cyclones depends largely on latitude. Closer to the tropics (e.g., the U.S. Gulf coast and southern U.S. Atlantic coast) tropical cyclones have a higher frequency of occurrence than farther poleward. Farther poleward, (e.g., in the northern U.S. Atlantic coast) extratropical cyclones and associated fronts occur more frequently and typically exhibit greater magnitudes. In the lower mid-latitude climates where both may occur, an issue of some debate concerns the relative importance of these storms, as it is unknown whether the higher frequency-lower magnitude extratropical cyclones have a greater geomorphic impact than the lower frequency-higher magnitude tropical cyclones.

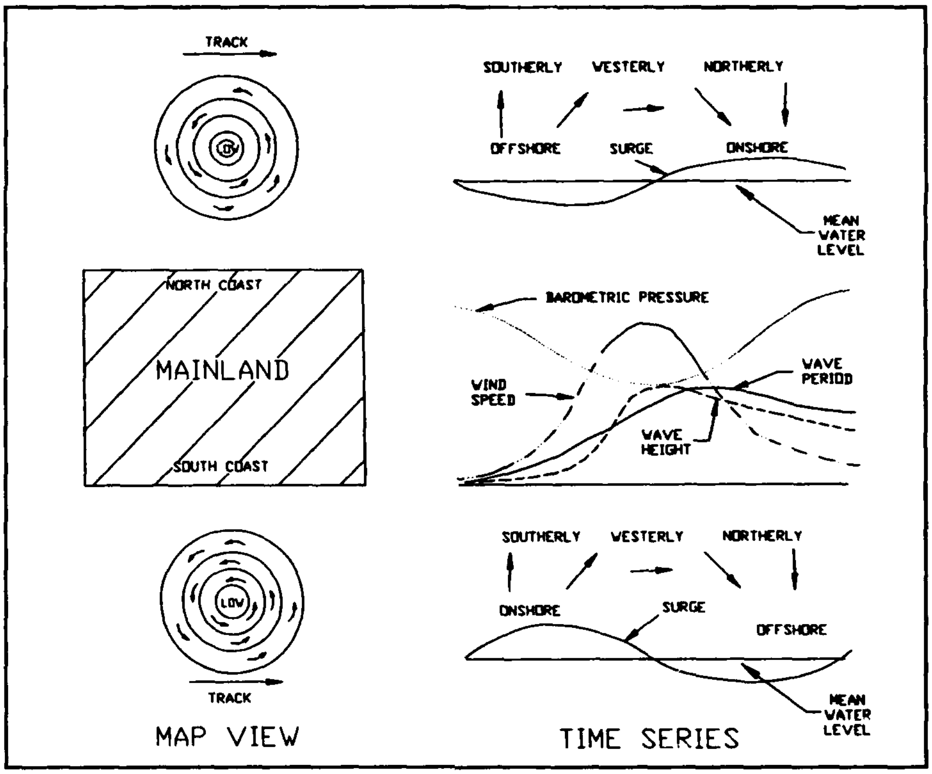


Figure 1. Changes in barometric pressure, wind speed, wind direction, wave height, wave period, and surge as a cyclone passes laterally offshore along a north- and south-facing coastline in the Northern Hemisphere (adapted from Carter 1988)

Waves

Waves are of great importance in the formation and variability of coastal geomorphic features. They are especially effective in the shore and shoreface areas, where they are capable of moving sediment directly or by generation of longshore currents. Wind waves, the most common type, are generated by wind stress on the water surface. The height and energy of wind waves increase with increasing wind velocity, fetch (the extent of water over which the wind blows), and wind duration.

Waves impinging on a coast are not necessarily produced by local winds in coastal waters. They may be produced by distant weather systems and reach the coast after traveling some distance, up to thousands of kilometers, outside the area in which they were generated. Such waves are called swells, while waves still within the area of generation are called seas. Swells have a longer period, or time of passage between two successive wave crests, and move faster than locally generated waves. Swells undergo a decay process in which the height becomes progressively less with distance traveled, although they

still retain most of their energy, and the wave period becomes longer. Two sets of long-period waves approaching the coast with slightly differing wavelengths may produce a resultant pattern, known as surf beat, that displays a periodic variation in height. Waves change character as they enter progressively shallower water, causing a variety of wave and breaker types, as well as several distinctive zones of wave action (Figure 2).

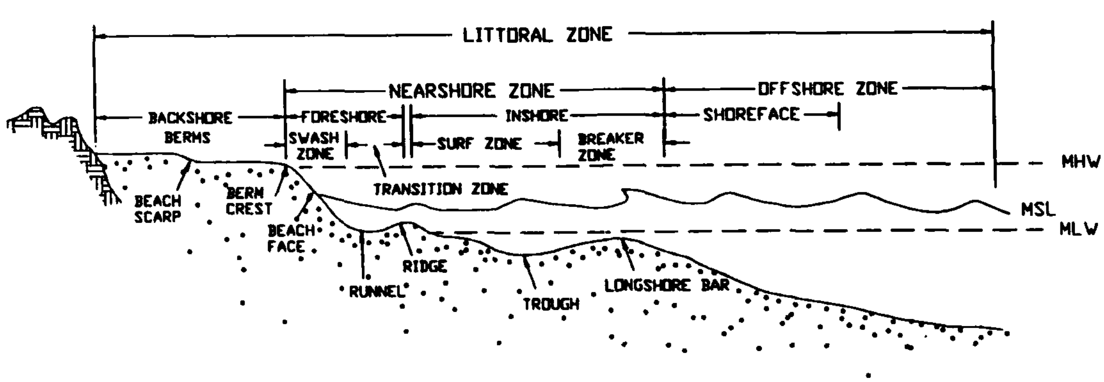


Figure 2. Distinctive zones of wave action and morphologic characteristics of the littoral zone

Waves approaching a coast undergo a transformation of certain characteristics which is called shoaling, because they have progressed into water depths in which frictional interaction with the bottom affects their motion. Significant transformation begins to occur when a wave enters water depths equal to about one half the wave length. This transformation causes the wave speed (celerity) to decrease, the wavelength to decrease, and the wave height to increase, since more energy is packed into a smaller area. The wave front may bend or change in direction as waves approach a shoreline and are affected by bottom topography. This phenomena is known as wave refraction.

Because of variations in bathymetry from place to place, the transformation of a given deepwater wave is site-specific and must be calculated on the basis of local bathymetric characteristics, bottom friction, and coastal configuration. Wave refraction bends the wave front, so that it becomes more nearly parallel with the bottom contours with decreasing water depths. For this reason orthogonals generally converge on headlands and diverge in bays (Figure 3). However, when waves cross irregular bathymetry, the waves may be refracted upon themselves.

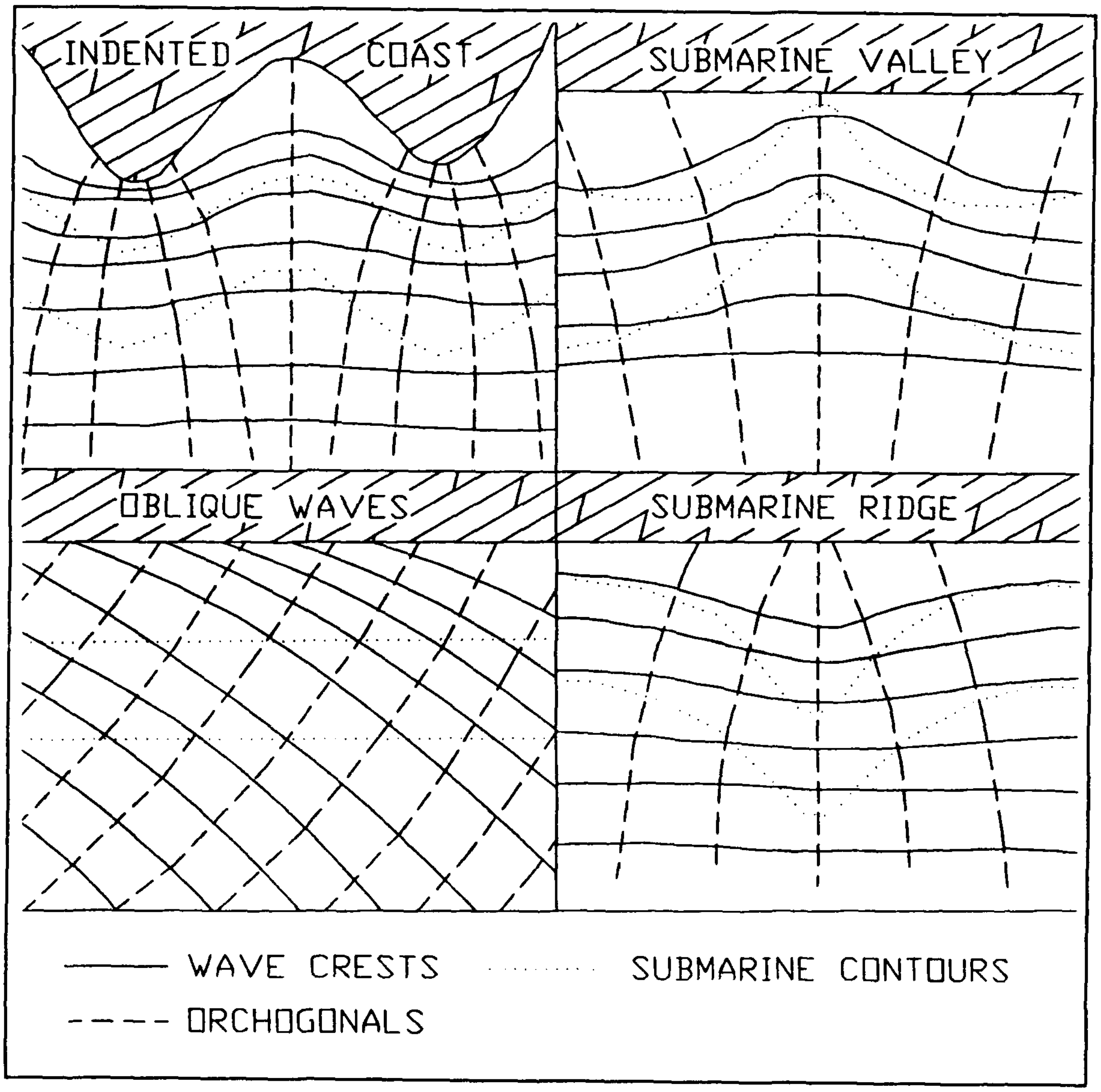


Figure 3. Wave refraction surrounding headlands and bays and over submarine morphologic features

As waves continue to enter shallower water depths, they may become unstable and break, becoming substantially distorted in shape from their deepwater form. Breaking waves follow a continuum that has been classified by Galvin (1968) into spilling, plunging, collapsing, and surging breakers. In general, steep deepwater waves and onshore winds approaching gentle slopes produce spilling breakers, where water spills down the steepened shoreward wave face (Figure 4). Plunging breakers are associated with long, low deepwater swells and intermediate slopes. Flatter waves breaking on a steeply sloping shore surge or collapse, with the base of the wave surging forward and the crest collapsing or disappearing.

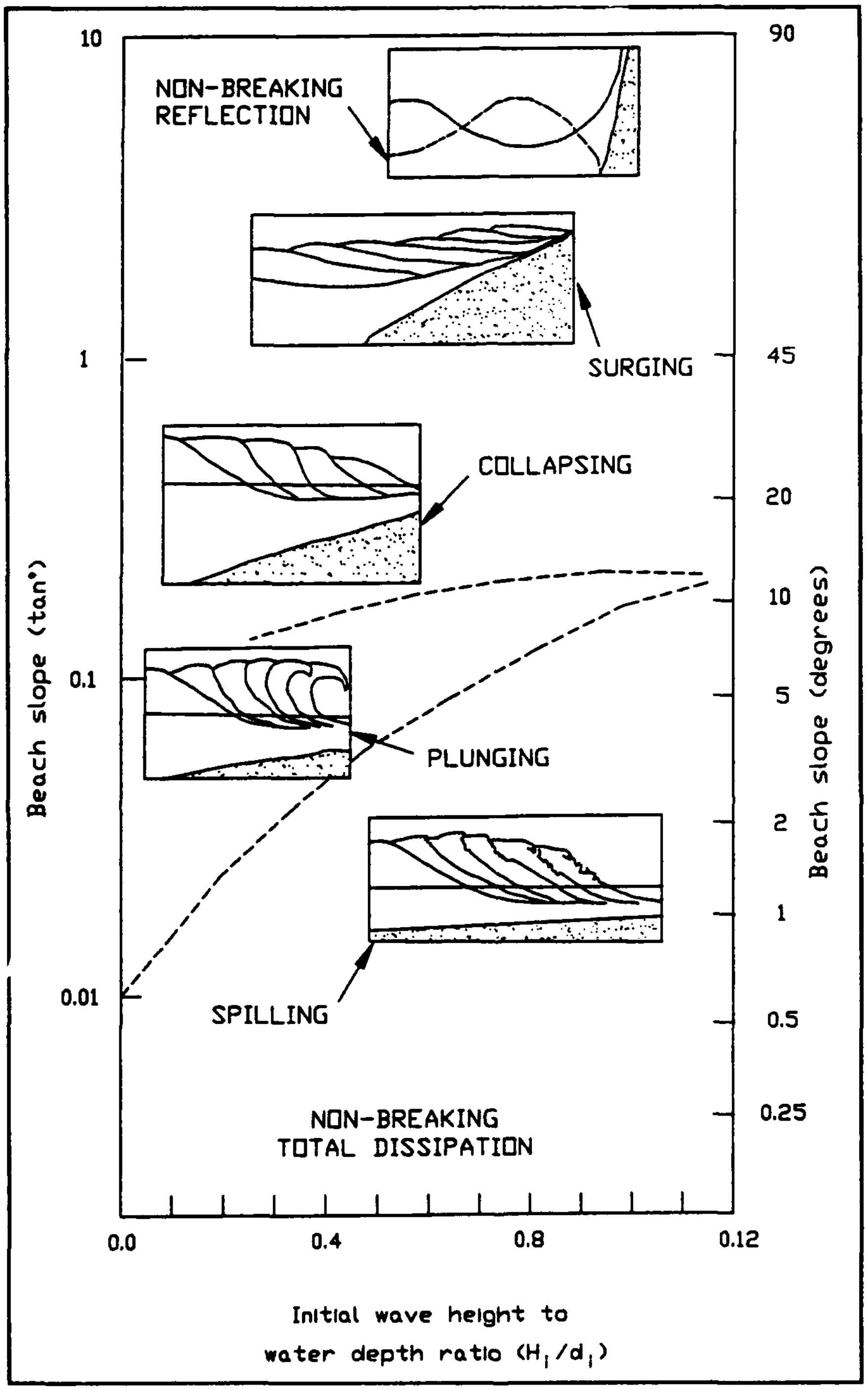


Figure 4. Appearance and classification of breaking waves according to Galvin (1968) combined with beach slope and wave height to water depth ratio as discussed by Street and Camfield (1966) (adapted from Carter 1988)

Once the waves break, a surf zone is generated in which much smaller waves are projected toward the foreshore and beach face. Upon breaking, the motion of the water particles in a wave changes from orbital to linear. The presence and the width of a surf zone are controlled by beach slope and tidal phase (Ingle 1966). Beaches with gentle slopes, which are typically composed of fine sand, are characterized by wide surf zones, whereas beaches with steep slopes, which are composed of shingle and cobble, often lack surf zones. Beaches of intermediate slope may have a surf zone at low tide, when wave action is over flatter portions of the beach profile, but may lack a surf zone at high tide when the waves break closer to shore over the steeper beach face.

The landward component of this linear motion, which surges onshore at a high velocity, is known as swash, while the lower-velocity return flow, which is driven by gravity, is known as backwash. As waves approach the beach at an angle, the oblique upward rush of swash, succeeded by the return of backwash down the beach, results in a longshore movement of sediment.

Deepwater wave characteristics can be estimated by an analysis of weather map data over the period of interest, a process known as hindcasting (Coastal Engineering Research Center (CERC) 1984). Waves may be observed from shipboard or shore, or can be measured by gages deployed on buoys, structures, or on the sea bottom. Refraction analyses can be performed manually by the orthogonal or wave-front method, or by computer methods when the offshore relief, wave approach direction, and wave period are known.

In addition to energy from incident waves, some energy may be transferred to secondary wave motions called edge waves, which develop at right angles to the shoreline. Edge waves have a maximum amplitude at the shoreline and decline seaward. They develop differently on shorelines of differing gradients. Many coastal forms and processes have been linked to edge wave characteristics (Holman 1983; Carter 1988).

The overall wave field at a site is typically formed by a combination of waves from several different sources, some local and some distant. Longer waves associated with storms reach deeper parts of the shoreface profile than do shorter waves associated with calmer conditions. In order to be useful in understanding the development and modification of coastal features and effects on coastal engineering works, data collected by the methods described above must cover a sufficient time span to record seasonal cyclic patterns and the occurrence of intense storms.

Tides

Tides are characterized by a rhythmic rise and fall, or flood and ebb, of sea level over a period of several hours. Because tides are generated by the gravitational forces associated with the moon and the sun, changes in the

declination of the moon and sun with the earth, and the relative position of these three astronomical bodies, influence the pattern and magnitude of tides. The timing and magnitude of high tide and low tide at any given location can be closely predicted by changes in these and other astronomical constituents.

Three major types of tides can be distinguished, based upon the pattern and frequency of occurrence of high and low tides during a tidal day. Tidal bulges form on both sides of the earth in response to a balance of forces associated with gravity, centripetal acceleration, and gravitational attraction (Komar 1976). The earth rotating on its axis typically produces semidiurnal tides, with two tidal maxima and minima every 24 hr and 50 min. This period exceeds the length of a solar day because of the advance of the moon's orbit. Diurnal tides display, on the average, one maximum and one minimum each day. Diurnal tides occur where the typical cycle is complicated by other astronomical factors. A combination of these characteristics, where two high and low tides having large diurnal inequality occur daily, are known as mixed tides. Mixed tides are commonly classified as being predominantly semi-diurnal or diurnal (Defant 1958). Important tide-generating constituents, expressed as a ratio of the major components influencing diurnal tides to those influencing semidiurnal tides, are used to measure these variations.

In general, tidal range is highest when the moon is full or new, and when the moon, sun, and earth are in alignment, or syzygy. These alignments produce the spring tides, in which the high tide is higher and the low tide is lower than average. Tidal range is lowest during the first and third quarters, when the moon and sun align perpendicularly with the earth, a condition known as quadrature. These less-pronounced high and low tides are collectively known as neap tides. The sequence from spring to neap to spring takes about 28 days, or one lunar month. Both spring and neap tides are affected by drag forces, which cause tidal range to lag syzygy and quadrature by up to 1.5 days.

Longer period cycles are generated by varying lunar and solar distances, caused by the ellipticity of both the moon's orbit around the earth, the earth's orbit around the sun, and the sun's declination. The moon's orbit around the earth is by far the more important of these. From perigee, when the moon is closest to the earth, to apogee, when it is farthest, represents a change in distance of about 13 percent. The earth is closer to the sun in early January by about 3 percent than its farthest position in early July, causing tidal range to be slightly greater in the Northern Hemisphere in the winter and fall than in the spring and summer.

Tidal range or magnitude is strongly affected by the depths and configuration of the land masses encountered. Davies (1964) introduced coastal classification divisions based on the spring-tide range, namely microtidal (< 2 m), mesotidal (2-4 m), and macrotidal (> 4 m). Tidal magnitude is influenced by the depths and configuration of the land masses encountered in crossing the continental shelves, which cause local resonances and reflections at land boundaries.

The type and frequency of occurrence of a wide variety of coastal landforms have been related to the tidal range (Hayes 1980) (Figure 5). Microtidal ranges occur on open ocean coasts and in certain landlocked seas. They are conducive to the development of river deltas, barrier islands, and spits. The Gulf of Mexico is an example of a microtidal environment. Macrotidal ranges occur where the tide is dissipated across a wide, shoaling slope, or confined to estuaries and gulfs whose land boundaries cause local resonances and reflections. They are conducive to the development of funnel-shaped estuaries, mudflats, and salt marshes. Locations with mesotidal ranges include features found in both microtidal and macrotidal environments, but are also well-known for well-developed tidal deltas.

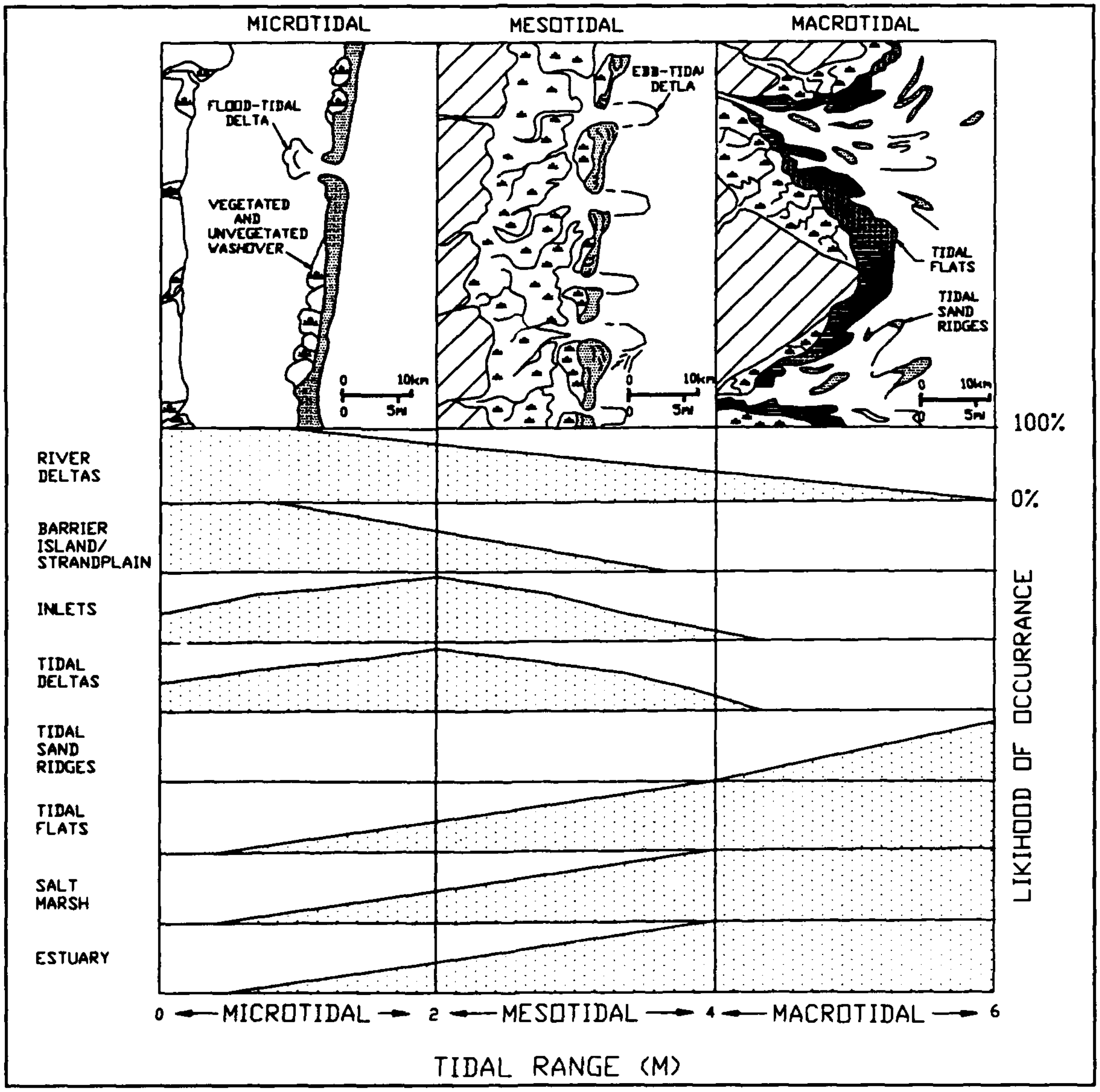


Figure 5. Macroscale morphology of microtidal, mesotidal, and macrotidal coastlines (modified from Hayes 1980)

Because tides are influenced by many factors and show great temporal and spatial variation, they play an important role in the variations of both processes and landforms in the coastal zone. The length of the drying period of the intertidal zone, changes in the water table of beaches, and the intensity and duration of tidal currents are all influenced by the type and magnitude of tides. Tides also influence the timing of morphologic changes, since cliff erosion and beach building may occur only intermittently at high tide. Long-term tidal variations, including solar semi-annual and 18.6-year components, may also be related to cycles of progradation and stabilization by mangroves on fluid mudflat coasts (Wells and Coleman 1981b).

Several landforms showing geomorphic variability are strongly influenced by tides. Tidal flats are depositional surfaces, alternately flooded and exposed. The formation, dimensions, and spacing of inlets are related to the total volume of water or tidal prism. Flood and ebb deltas, as well as associated features at inlets, are controlled by tidal current hydraulics. Tidal channels of various types, including intertidal channels, tidal creeks, and tidal rivers, are influenced to differing degrees by tides.

Currents

Near the coast, currents are effective agents of erosion and transport of sediments and, consequently, have an impact on the geomorphic variability of coasts. Mechanisms responsible for the generation of coastal currents include winds, waves, and tides.

Wind-driven currents, in combination with density currents, comprise much of the large-scale oceanic circulation, as well as much of the circulation in marginal seas. These currents, generated by the drag of wind on the ocean surface, are deflected up to 45 deg as a result of the Coriolis effect, although appreciably less in shallower waters. Ocean circulation following the parallels is predominantly wind-driven, whereas circulation following the meridians is largely density-driven (Mosetti 1982). Wind-driven currents show greater influence on geomorphic variability in marginal seas than oceans, because sediment transport and beach modification are facilitated by shallower depths.

Wave-generated currents can influence any part of the shore, shoreface, and Continental Shelf; however, the most important currents for coastal development are longshore currents. Longshore currents are generated between the breaker zone and shore by waves approaching the shore at an angle. The wave height, period, direction of approach, and bottom slope are important in determining the direction and characteristics of longshore currents. Flow velocities of longshore currents are typically around 30 cm/sec, but can be substantially higher. While such velocities are often unable to entrain materials, suspension by concurrent waves may greatly increase the effectiveness of sediment transport.

In some places, the hydraulic head of water carried onshore by waves may return across the surf zone as a concentrated jet, known as a rip current (Bowen 1969; Bowen and Inman 1969). Rip currents are relatively narrow and strong, capable of moving sediment seaward through the breakers and the inshore bar field. Although near-bottom flow usually dissipates rapidly, surface flow may persist for greater distances. The geologic signature of rip currents can be seen on the seafloor well seaward of the surf zone (Morang and McMaster 1980). Rip currents tend to be found at positions of lowest breaker height.

The existence of rip currents, and the manner in which they are fed, depends upon whether the nearshore current system is dominated by a cell circulation system or a longshore current. Typically, both situations are present simultaneously (Shepard and Inman 1950). In the cell circulation system, the rip currents are fed by a series of longshore currents that increase in velocity from a minimum midway between two adjacent rips to a maximum just before turning seaward into the rip (Figure 6). If the wave approach angle is large, at least 5 deg to 10 deg, a strong longshore current will be generated that is continuous along the shoreline. The intermediate condition with a smaller breaker angle is known as general circulation. General circulation may result in an asymmetrical current pattern, with velocities at their minimum level updrift from the rip current and increasing to a maximum before turning seaward to the next rip current (Komar and Inman 1970).

Tidal currents are horizontal water movements caused by the rhythmic rise and fall of the sea surface. Their magnitude generally increases with increasing tidal range. Tidal currents force water to move both at the surface and at depth, and generally have the greatest velocities during the flood and ebb, at mid-stage between high tide and low tide. In the open ocean, tidal current velocities rarely exceed 1 m/sec. In restricted passages and shallow coastal shelves, however, velocities are greater and can exceed 4 m/sec in some channels.

Storm Surges

Sustained and strong wind stresses on a water body, often developed under tropical and extratropical storm systems, not only create waves but also produce a horizontal flow of water in the general direction of the wind. When this flow approaches a coast and is affected by shoaling, there is a sustained increase in the water levels, known as a storm surge. Storm surges may be as much as 3 to 6 m during major hurricanes or extreme seiches on the Great Lakes, and may be augmented by decreases in atmospheric pressure. This anomalously high water level can cause flooding of the beach, dunes, and

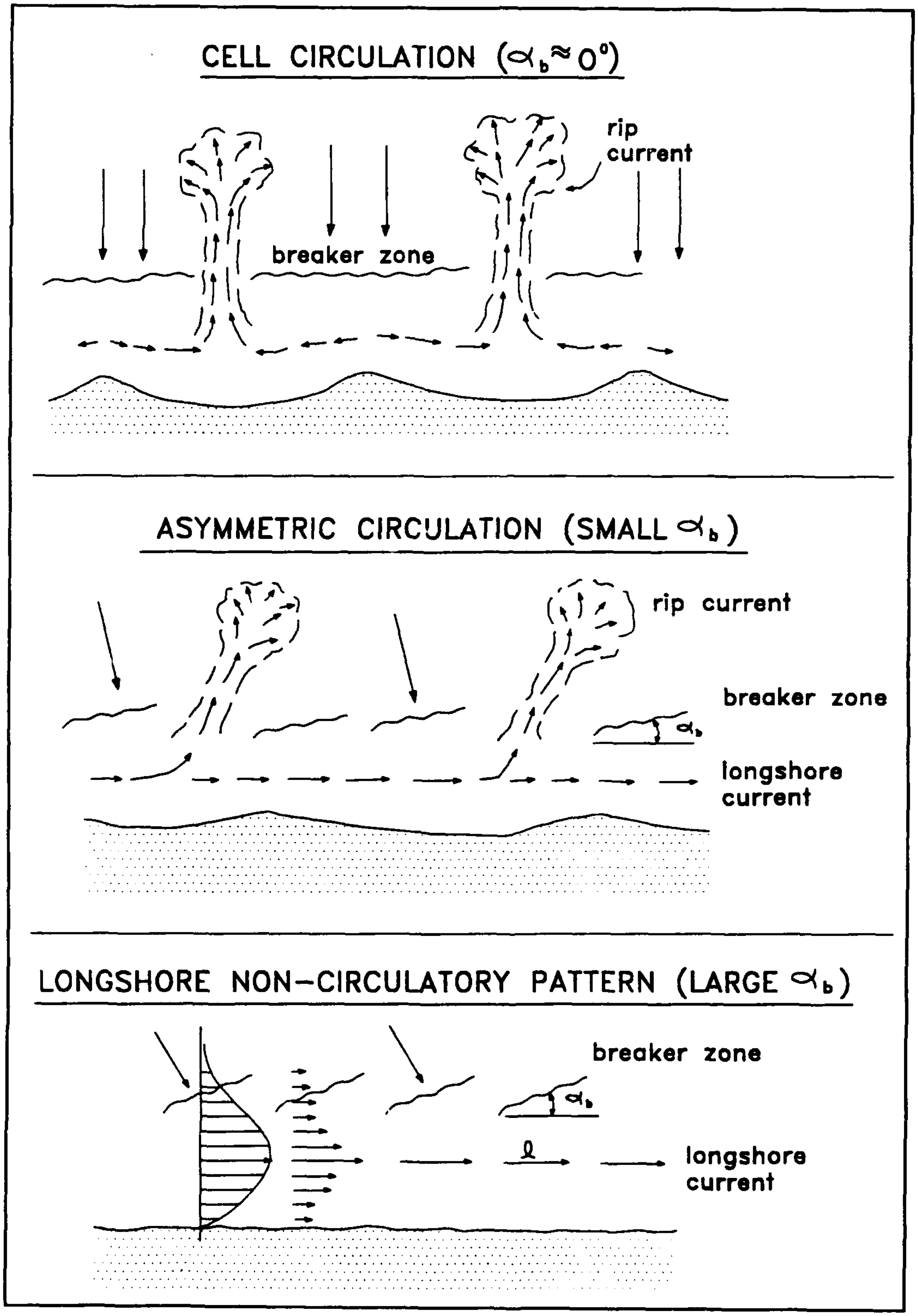


Figure 6. Wave-induced nearshore currents as determined by breaker angle. Cell circulation pattern with well-developed rip currents occurs when the breaker angle is close to zero. Asymmetric circulation with longshore currents feeds rip currents when breaker angle is small. A longshore non-circulatory pattern occurs with an oblique wave approach and large breaker angle

inland areas, can extend the zone of wave attack inshore, and can increase hazards and damage. Strong offshore winds can similarly produce decreased water levels.

Storm surges, as products of the weather, can be studied by hindcast methods from historical weather charts in the same general manner as waves. The magnitude of water-level change is dependent upon wind velocity and direction, fetch, depth of water, and slope of the inner shelf (*Shore Protection Manual* 1984), as well as the orientation of the storm system in relation to the coast. As with tides, the configuration of the coast may influence water levels, with increased elevations being particularly common in embayments.

In many cases more direct records of storm surge can be obtained from tide gages and measurement of high-water marks following the storm. In recent years, there has been an increasing volume of work in storm surge modelling and prediction. The Sea, Land and Overland Surges from Hurricanes (SLOSH) model (Jelesnianski and Chen 1982) and the Special Program to List the Amplitudes of Surge from Hurricanes (SPLASH) models (Jelesnianski 1972) are among the most widely used.

Tsunamis

Tsunami is a Japanese term used widely in the scientific literature to refer to a train of progressive long waves generated by an impulsive disturbance either in the ocean or a small body of water. A tsunami is usually caused by a severe earthquake. Tsunami waves have periods of several minutes to several hours and can move at speeds of 700 to 900 km/hr in the open ocean, depending upon water depth. Upon approaching shallow water, their speed decreases but their wave height increases dramatically.

While tsunamis are rare, they have the potential to cause extreme geomorphic change and damage to structures. Crustal movements around the Pacific basin are likely to cause tsunamis. Islands in the center of the Pacific basin, such as Hawaii, can receive effects from multiple sources, and are particularly vulnerable.

Relative Sea Level

Changes in relative sea level occur when there is (a) a general eustatic variation of sea level, (b) uplift or subsidence of a coastal landmass, or (c) a combination of the two. The net gain or retreat of land along coasts is influenced by the combination of sea level changes and changes in erosion or accretion over time (Bloom 1965). Transgression, the landward migration of the shoreline, usually occurs with rising relative sea level unless a high rate of

deposition can offset this tendency (Curray 1964). Regression, the seaward migration of the shoreline, typically occurs with falling relative sea level, unless high rates of erosion offset this tendency (Payton 1977).

Eustatic changes are primarily the result of changes in the volume of water in the oceans or changes in the dimensions of ocean basins. Although there are many reasons for variations in ocean volume, the growth, or waxing, and decay, or waning, of large continental and alpine glaciers have been the primary causes of fluctuations in the past 2 million years (the Pleistocene and Holocene Epochs) (Emery and Aubrey 1991). During glacial episodes, enormous amounts of water were locked up in the massive continental and widespread alpine glaciers, causing a substantial reduction in the volume of oceanic water and a reduction of sea level to more than 100 m below the present level. During interglacial episodes, the amount of meltwater that returned to the ocean caused a rise in sea level that at times exceeded the present elevation by more than 80 m.

The last major glaciation, known as the late Wisconsin age in North America, reduced sea level to over 100 m below the present. The waning of that glaciation precipitated a rapid rise of sea level commencing about 18,000 years ago, reaching near the present level about 4,000 years ago, when sea level became more or less stable. Because large glaciers still exist in Antarctica, Greenland, and some mountainous areas, the present sea level is below that of some prior interglacial periods when a more extensive melting of glacial ice occurred.

The fluctuations of sea level during the Pleistocene and Holocene Epochs have caused the shoreline to migrate back and forth, alternately exposing and submerging large parts of the present continental shelves and areas inland of the present shoreline. These transgressive and regressive events have left depositional features of terrestrial origin in marine environments, and features of marine origin in terrestrial environments.

Studies of historical tide records in recent years indicate that a rise in sea level is taking place in many locations (Hicks 1978; 1983; Gornitz, Lebedeff, and Hansen 1982). This rising trend, which some scientists believe may accelerate in the future, has been attributed by some to the general warming of the planet accompanied by glacial melting, possibly caused by increased concentrations of carbon dioxide (CO_2), methane (CH_4), and other gases in the earth's atmosphere which trap longwave-electromagnetic radiation. Projection of future trends based on past sea level history is difficult because of the relatively short time that widespread accurate data have been systematically collected, because of problems in correcting for vertical earth movements, and because of difficulties in projecting trends.

Local and regional changes in relative sea level also occur where coastal landmasses are uplifted or depressed as a result of tectonic activity, sediment compaction, or unloading and loading by glacial ice or water. These movements may increase or decrease the effects of eustatic sea level fluctuations,

and may result in local horizontal transgressions or regressions of sea level or vertical submergence or uplifting of terraces.

Tectonic movements may be so slow that they have little influence on geomorphic variability. On the other hand, they may occur as dramatically and suddenly as in the Great Alaskan Earthquake of 1964, which caused upward displacement of 10 m in places, and downward displacement of 2 m in others, over a period of only a few months (Hicks 1972). In general, these movements tend to be greatest at margins of tectonic plates and, in particular, at converging plates.

Consolidation and dewatering, which cause a reduction in volume, are particularly prominent in areas which experience rapid sedimentation, such as the Mississippi Delta of southeastern Louisiana. Surface and subsurface sediments, which are poorly consolidated and include organic-rich materials, may experience very high rates of compaction. In general, the rates of sediment consolidation decrease with increasing depth. They also decrease with increasing age, especially in environments of recent sedimentation.

Unloading and loading of the lithosphere by glacial ice and water may also produce vertical and horizontal movements, further complicating the history of relative sea level changes along coastlines. Isostatic depression of the lithosphere, which occurs with glacial advance over the continents, is proportional to the ratio between the density of ice and that of the mantle (Andrews 1974). Isostatic or crustal rebound follows deglaciation. Rates of uplift, estimated from raised shorelines, may exceed 20 mm/yr (Smith and Dawson 1983). Although much slower rates of isostatic compensation are attributed to erosional unloading, these rates may be sustained over longer periods.

Lithology and Weathering

The nature of rocks and sediment deposits is of great importance in determining the inherited morphology and the development and modification of coastal features. The lithologic factor having the greatest effect on coastal features is the degree of consolidation. This governs the ability of rock material to be eroded and transported, which, in turn, influences the form and stability of geomorphic elements occurring in the coastal zone.

Consolidated rock coastlines are often hilly or mountainous, with the exception of reefs, which are generally low-lying. The inherited morphology is usually prominent, with erosional features being more numerous than depositional features. Some geomorphic variability is attributable to rock type, interbedding, jointing, and dip and strike of strata (Figure 7). Temporal geomorphic variability is not as great as on unconsolidated coasts because consolidated rocks are highly resistant. However, in some places, rocky coasts may be modified by depositional features. These include locations where

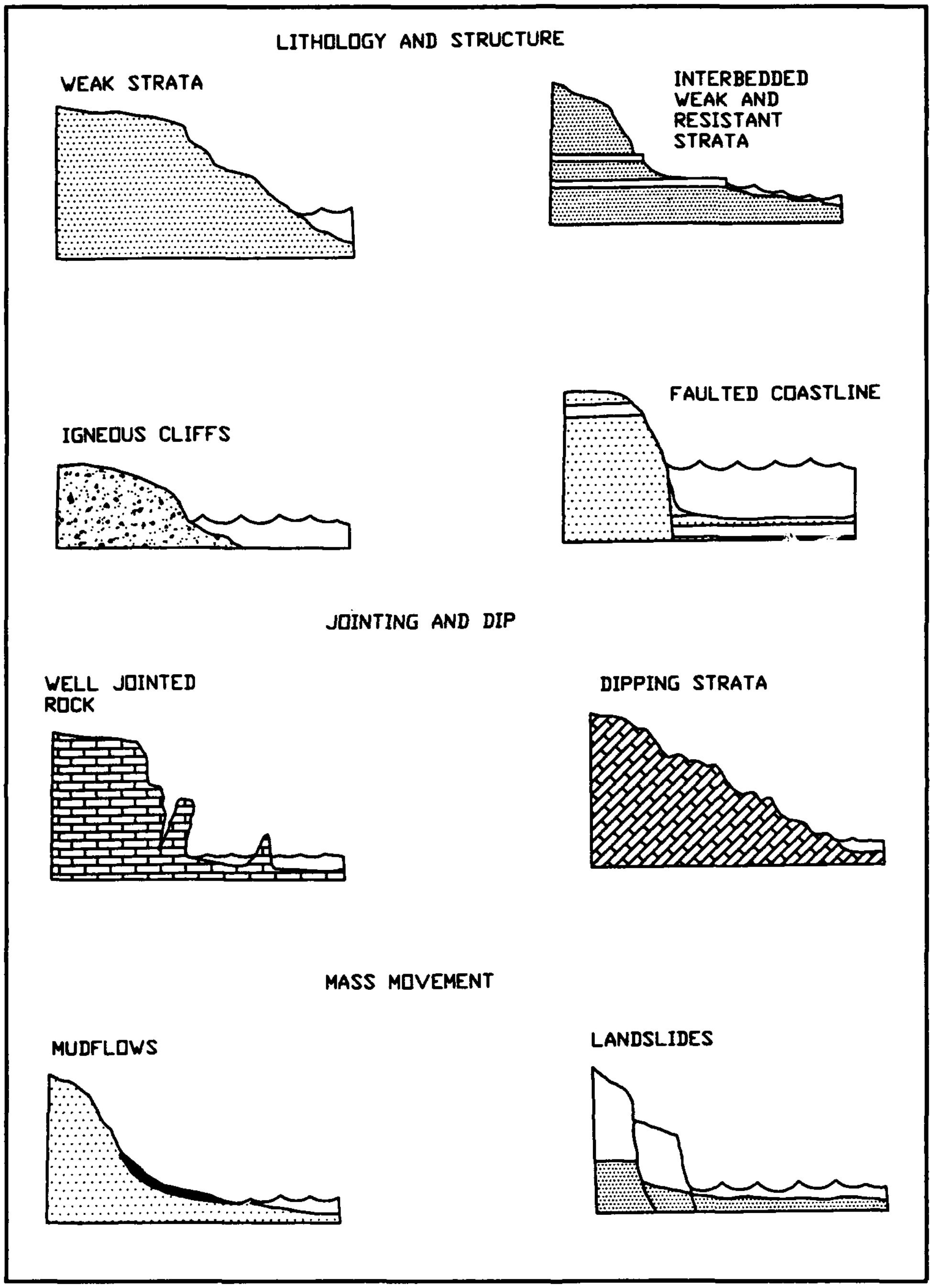


Figure 7. Some aspects of geomorphic variability attributable to lithology, structure, and mass movement along semi-consolidated and consolidated coasts in cross section

rivers deliver sufficient quantities of sediment for the construction of extensive spits and baymouth barriers, or where erosion of cliffs or reefs exceeds transport of sediments by marine processes.

Unconsolidated or poorly consolidated materials are found in low-lying coastal plains and deltaic complexes. Along coasts of unconsolidated materials, large amounts of sediment typically are available and morphologic variations can occur rapidly. Relict geomorphic features are readily altered in this environment. Depositional features are likely to be more numerous than erosional features. A detailed description of coasts composed of unconsolidated sediments follows in a section on sediment character.

Both lithology and jointing influence the resistance of a rocky coastline to weathering and erosion. Additional factors are susceptibility to weathering (mechanical or chemical), hardness of the constituent minerals and cementation, the nature and density of voids, and climatic conditions. Mechanical weathering involves the breakdown or disintegration of rock without any substantial degree of chemical change. Examples are processes such as temperature change, crystallization by salt or frost, wetting and drying, changes in overburden, and organic activities. Chemical weathering involves the decomposition or decay of minerals because of hydration and hydrolysis, oxidation and reduction, solution and carbonation, chelation, and/or biochemical changes.

In a given area, weathering or erosion may occur at different rates. Various rock types may be more or less susceptible to erosion by waves, tides, and currents. As a result, differential weathering and erosion may produce uneven coastlines where headlands are formed of resistant materials and bays occur in less resistant materials (Shepard and Grant 1947) (Figure 8). Small-scale effects of differential weathering cause rocks to have uneven pitting or surface characteristics. Coastal configuration at differing scales may also be influenced by terrestrial agents that cause erosion, such as running water, ice, wind, and groundwater.

Organisms

Marine organisms may play either a destructive or constructive role in the formation of coastal sediment deposits. Some organisms, including species of algae, mollusks, echinoids, worms, and sponges have the ability to bore into rock for protection against predators or to obtain anchorage on the bottom. The resulting weakening and breakdown of the rocks makes them more susceptible to wave erosion. Physical and chemical weathering generally break down and dissolve organically produced or biogenic materials more rapidly than clastic sediments.

Biological activity also plays a constructive role in the formation of coastal sediment deposits. Although biogenic materials are more easily destroyed than terrestrial clastics, new sources are continually being provided. Many organisms that inhabit the submerged part of the coastal zone contain hard parts, usually composed of calcium carbonate. Upon the death of the

Figure 8. Aerial photograph of Pta. de la Garita, Cabo de la Aguja, Columbia (Feb 1981). A pocket beach has formed between two resistant headlands. To the right of the headlands, waves approach the coastline at a steep angle, causing longshore currents which transport sediments away from the headland. Photograph taken from an altitude of 600 m; distance between the headlands is about 0.5 km.

organism, the hard materials become part of the bottom sediments. Mollusks, calcareous algae, barnacles, bryozoa, and foraminifera are important elements of coastal sediment deposits in many places.

In most places, mollusks are the principal organic shell contribution to coastal sediments. Breakdown of the larger shells into sand and granule-size material usually occurs where shells are exposed to boring organisms and the action of waves and currents. Some organisms have segmented shells that separate into smaller particles soon after death. Common examples of this are certain calcareous algae such as *Halimeda* sp. and barnacles. Both of these organisms are made up of numerous parts, which are composed of calcium carbonate.

In areas of high biological production and/or low input of terrestrial sediment, biogenic sediment particles may exceed the inorganic particles in number. In certain environments, such as coral reef areas, biogenic material is dominant, and sometimes the sole material type. Accumulation of biogenic material in sediments seems to be more important in offshore areas than in beach and dune sediment, presumably because of the higher destruction rate of

shell material in the turbulent beach environment, especially where the shell material is mixed with substantial amounts of quartz sand.

It is not uncommon to encounter shell material that has been produced in an earlier time under different environmental circumstances than the present (relict material). From North Carolina to Florida, relict material may be a significant element in coastal deposits. One way in which coastal deposits may be preserved is by secondary diagenetic changes. An example is the formation of beach rock, which involves the consolidation of beach sand by interstitial cement composed chiefly of calcium carbonate (Higgins 1968). For example, the Anastasia Formation beach rock, a coquina limestone, may outcrop in the surf zone as nearshore reefs or as a low cliff in the berm area (Figure 9).

In addition to contributing sediment particles, several organisms are capable of building reefs on the sea floor. Corals are the best known reef builders and usually produce the largest reef structures. However, various types of worms, mollusks, bryozoa, and coralline algae are also capable of reef construction.

Figure 9. Outcrops of Pleistocene coquina rock of the Anastasia Formation on a beach in East Florida. Erosion of such rocks contributes sediment particles to the beach

Coral reefs are best developed in subtropical and tropical climates but certain species of coral can tolerate colder water temperatures. The actual amount of coral in a reef varies. It can be relatively small, providing a framework that fills out with the hard parts of plants and animals that flourish in the typical coral reef environment. Coralline algae are important in reef construction because they form calcium carbonate crusts that help hold the different reef materials together.

Mass Wasting and Mass Transport

Once clastic and biogenic materials are broken apart by mechanical and chemical weathering, they are more susceptible to downslope movement under the influence of gravitational processes. These processes, collectively termed mass wasting, may vary in scale, occurring either slowly or quickly in a number of parent materials. Causes and mechanics of mass wasting are variable, but depend in part on slope, soil moisture, and physical properties. While most attention is given to subaerial mass wasting, submarine failures occur on deltas and on the Continental Shelf (Prior and Coleman 1978).

The shape of failures and distribution of debris may take on a variety of forms. Thus, in coastal settings, mass wasting may directly influence geomorphic variability (Figure 7). Along cliff coasts, especially in unconsolidated material, gravity movement is important and is often aided by wave undercutting of the base. Waves and currents are important in removing mass wasting debris, thus reexposing the cliff face to wave attack (Figure 8).

Mass wasting of hillslopes, whether at the coast or inland, may also facilitate sediment supply. Mass wasting directly by gravity is distinguished from other gravity-induced movements in which the material is carried by transporting agents such as running water, groundwater, ice, snow, and air. Direct movement by these agents is termed mass transport, although, in nature, mass wasting and mass transport merge into each other so that in some cases the distinction becomes arbitrary. Abrasion or mechanical wear of materials by solid particles transported in fluids, and corrosion or chemical wear from the reaction of rocks with substances in water are additional processes involved in the erosion of materials.

Materials may be transported in one of three major modes within a fluid (dissolved load, suspended load, or bed load). The dissolved load consists of material transported in solution. The suspended load consists primarily of fine particles, which are entrained and maintained into the flow primarily by turbulent mixing processes. In bed load transport, particles move by rolling, sliding, or saltating at velocities less than those of the surrounding flow.

The great majority of material is transported from inland areas to coasts by running water. The magnitude of sediment yield by running water is con-

trolled by five main groups of factors: (a) precipitation and runoff characteristics; (b) soil resistance; (c) basin topography; (d) vegetation cover; and (e) land use. Of these, most consider mean annual precipitation or runoff to be the most important variable. Although studies show great disparity, the greatest sediment yields occur in semi-arid regions and very humid regions (Knighton 1984). Only a small proportion of the sediment mobilized from inland areas on a given occasion will reach the coast, while most is stored temporarily in the basin. There is large geographical variability in concentration and distribution of sediment carried from sources to the coast by running water.

In a fluid, the entrainment, transport, and deposition of sediment particles are often defined as functions of sediment diameter and mean fluid velocity. In flowing water, the Hjulstrom curve indicates that the threshold velocity is at a minimum for medium size quartz sand particles, and that higher velocities are necessary to entrain both finer and coarser sediments (Figure 10), for reasons to be discussed shortly. Greater velocities are required to transport materials of similar size in wind as opposed to water (Bagnold 1941). Velocity alone does not control the capacity of a fluid to entrain particles. Other characteristics include the viscosity of the fluid, the nature of the fluid motion, the character and shape of the bed materials, and the impact of saltating or bouncing grains.

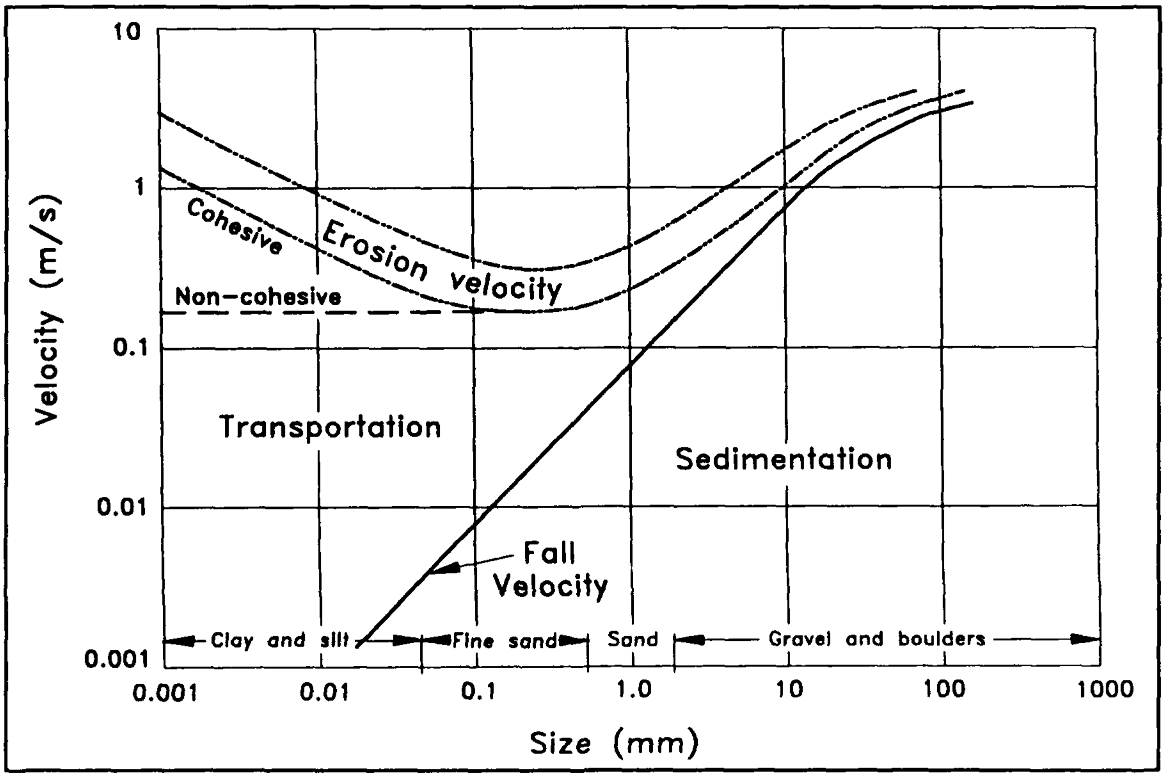


Figure 10. Conditions for erosion, transportation, and deposition of particles in water according to Hjulstrom (1935)

Fine, cohesive sediments respond differently to hydrodynamic forces than do non-cohesive sediments. Fine sediments are made up largely of clay minerals with an interlayered crystal structure, and normally have a negative surface charge. Cohesion results from inter-particle surface attractions between clay minerals, which are promoted in sea water, and then reinforced by organic secretions. Thus, for noncohesive sediments, the main stabilizing force is the particle weight, whereas cohesive sediments are stabilized by interparticle adhesion and organic binding.

The relative importance of running water, wind, ice, or groundwater in transporting materials from inland areas to the coast varies greatly with local conditions. Running water is most important for transporting solid and dissolved materials where large rivers meet the coast. Locally, smaller rivers may be important, and reworking of ancestral river deposits is also often significant. As with running water, groundwater may be important for transportation of solutes or dissolved materials. Wind is important in arid areas, although not necessarily more important than running water. Transportation by ice is much slower than by water and wind, so that ice is probably only important over long-term periods at high latitudes.

Between the offshore and coastal zones, mass transport may take place by a variety of mechanisms. Erosion and transport of offshore sediments are chiefly accomplished by waves and wave-generated littoral currents in the shore and upper shoreface areas. Considerable difficulties arise when attempting to apply sediment movement thresholds in oscillatory flow, as the water particles under waves reverse their direction of flow and accelerate and decelerate under each pulse. The threshold condition for movement appears to be better related to the ratio between grain diameter and orbital diameter of the water particles (Komar and Miller 1973).

Storm-generated waves and currents can effectively erode and transport material in deeper waters of the continental shelf, and it is likely that they play a role in modifying and moving large sediment features, such as shelf shoals. Mass transport has both a longshore component, parallel to the shore, and a cross-shore component, which may be onshore or offshore. Lower-energy events mostly transport sediment onshore, while larger events transport sediment offshore and sometimes onshore, if overwashing and overtopping occur.

Sediment Character

Unconsolidated coastal sediments may be composed of a variety of materials, which range in size and shape, mineralogy, density, and other properties. Clastic sediments are comprised of rock detrital grains, whereas biogenic sediments are comprised primarily of calcium carbonate grains from shells, skeletons, and invertebrates. The characteristics of these materials account for

much of the geomorphic variability of clastic shorelines and provide information that may assist in interpreting past processes or in predicting the transport potential of sediments.

Sedimentary particles show a range in size and are classified according to diameter, assuming the particles are roughly spheroid in shape. These sizes are divided into several major classes, in order of increasing diameter, being clay, silt, very fine to very coarse sand, granule, pebble, cobble, and boulder (Table 1). Generally, clay minerals are cohesive, being held together by electrolytic forces. Coarser sediments, which make up the bulk of coastal sediments, are considered noncohesive or cohesionless.

The type of sediments found in a location depends on the source and supply of materials and the energy of the environment. If the nature of the source and supply are held constant, coarser sediments are deposited in high-energy environments and produce beaches of steeper slope, whereas finer sediments are deposited in low-energy environments and are associated with beaches of gentler slope. Mixtures of differing populations of sediments are common, including combinations of coarse and fine, clastic and biogenic, and differing source regions. These contribute further to the geomorphic variability of coasts.

Sandy coastlines are predominant worldwide. Sand may be supplied by rivers, by adjacent parts of the coasts including beaches, headlands and cliffs, by offshore sources, or by wind. Silt and clay coastal sediments occur in generally lower energy settings such as lagoons and back barriers. Near large rivers (Wells and Coleman 1981a) and in glaciolacustrine settings, silts and clays may dominate coastal materials.

Pebble and cobble coastlines, called shingle beaches in Britain, are more common in areas of glacial and fluvioglacial sediments (Bird 1969; King 1982), in areas where coastal rock formations yield debris of appropriate size (Bird 1969), and in localized areas where rivers deliver coarse materials to the shore. Beaches of granule-sized particles are rare (King 1982).

Cobbles and pebbles on beaches have a variety of characteristic shapes, including disks, rods, and spheres, which are found in characteristic zones of the beach profile. The distribution of the various sediment shapes on beach profiles is controlled largely by selective sorting (Bluck 1967). Contrary to usual descriptions, marine abrasion appears not to be the predominant cause of the disk shape, in that the largest oblate disks are found near the high tide mark and thus are least worked by the sea.

Beaches of terrigenous origin generally have quartz as the most abundant mineral, accompanied by varying proportions of feldspar, mica, other light minerals, and heavy minerals. Quartz is dominant because it is the most abundant mineral in the earth's crust, as well as being mechanically durable and chemically inert (Jackson 1970). Pebble and cobble beaches have highly varied mineralogy, depending upon the nature of the source. Clay minerals

Table 1
Sediment Particle Sizes

Unified Soils Classification	ASTM Mesh No.	MM size	PHI Size	Wentworth Classification	
		4096.00	-12.0		
		1024.00	-10.0	Boulder	
		256.00	-8.0		G
Cobble		128.00	-7.0		
		107.64	-6.75	Cobble	
		90.51	-6.5		
		76.00	-6.25		R
		64.00	-6.0		
		58.82	-5.75		
		45.26	-5.5		
		38.00	-5.25		A
Coarse Gravel		32.00	-5.0		
		26.91	-4.75		
		22.63	-4.5		
		19.00	-4.25		V
		16.00	-4.0	Pebble	
		13.45	-3.75		
		11.31	-3.5		
		9.51	-3.25		E
Fine Gravel	2.5	8.00	-3.0		
	3	6.73	-2.75		
	3.5	5.66	-2.5		
	4	4.76	-2.25		L
	5	4.00	-2.0		
	6	3.36	-1.75		
Coarse Sand	7	2.85	-1.5	Granule	
	8	2.35	-1.25		
	10	2.00	-1.0		
	12	1.68	-0.75		
	14	1.41	-0.5	Very Coarse	
	16	1.19	-0.25		
Medium Sand	18	1.00	0.0		S
	20	0.84	0.25		
	25	0.71	0.5	Coarse	
	30	0.59	0.75		
	35	0.50	1.0		A
	40	0.42	1.25		
	45	0.35	1.5	Medium	
	50	0.30	1.75		
	60	0.25	2.0		N
	70	0.210	2.25		
Fine Sand	80	0.177	2.5	Fine	
	100	0.149	2.75		
	120	0.125	3.0		D
	140	0.105	3.25		
	170	0.088	3.5	Very Fine	
	200	0.074	3.75		
	230	0.0625	4.0		
	270	0.053	4.25		
	325	0.044	4.5		
Silt	400	0.037	4.75	Silt	M
		0.031	5.0		
		0.0156	6.0		
		0.0078	7.0		
		0.0039	8.0		U
		0.0020	9.0		
		0.00098	10.0	Clay	
Clay		0.00049	11.0		
		0.00024	12.0		D
		0.00012	13.0		
		0.00006	14.0	Colloid	

also have varied composition, the most common being kaolinite, smectite, illite, and chlorite.

Shell fragments, coral debris, algal material, and oolites are common on tropical coasts and islands, in temperate regions where shelly organisms are abundant offshore, or along coasts with calcareous cliffs. The mineralogy of the fragments is primarily calcareous, although dolomitic, phosphatic, and siliceous materials may be present. Most calcium carbonate grains are biogenic since they are made up largely of shells, skeletons, and invertebrates. In some tropical areas where the water is super-saturated with calcium carbonate, non-biogenic precipitates of calcium carbonate called oolites may form.

Along volcanic coasts, such as Hawaii and the north shore of Martinique, black sand beaches of basalt and pumice may be dominant. Driftwood and timber are common, particularly in the vicinity of lumbering areas, although they can be found on virtually any coast. Along cliff coasts, logs may facilitate erosion by abrasion. Other organic materials, such as peats, are common coastal materials. They are formed in low-energy areas such as vegetated marshes and swamps.

Characteristics of the sediments can provide clues to their origin and depositional environments. Size analysis can help distinguish beach, dune, and eolian flat environments (Mason and Folk 1958), and the shape, inclusions, and optical characteristics of a sediment may also assist in determining its geologic origin. The mineralogy of the materials and heavy minerals can be used as a tracer to indicate the source or provenance of sediments. Heavy minerals typically form a minor constituent of the original rock and have a specific gravity greater than quartz or feldspar, with a density of 2.8 gm/cu cm being a generally accepted lower limit (Brenninkmeyer 1978). Because of their greater density, heavy minerals respond differently to the sorting and concentration that occur in marine processes. The utility of various heavy minerals in determining provenance varies according to their occurrence (common to very rare) and their stability. Statistical techniques, including factor analysis (Clemens and Komar 1988; Komar et al. 1989), may be useful in determining dispersal patterns and sorting of heavy minerals.

Just as rocky coasts may produce significant zones of unconsolidated sediment because of erosion, coasts of unconsolidated sediments can develop lithified zones because of diagenetic changes. One of the most common secondary diagenetic changes is the formation of beach rock, involving the consolidation of beach sand with an interstitial cement composed chiefly of calcium carbonate (Higgins 1968). Beach rock develops best on tropical and subtropical coasts, although it has also been reported on temperate coasts. Most researchers think that the formation of beach rock takes place underground, near the top of the water-saturated zone of the beach. Thus, beach rock is generally not visible unless shoreline recession has occurred and the overlying sediment has been washed away.

Sediment Supply and Human Activity

Sediment supply at a given coastal segment can be a critical factor in the morphology of unconsolidated coastal features such as beaches, deltas, and capes. The volumetric accounting of the material lost or gained constitutes a sediment budget. Longshore transport out of or into an area is the chief cause of loss or gain along mainland coasts and most barrier beaches (Bowen and Inman 1966). Other important reasons for gains include river transport, sea cliff erosion, onshore transport, biogenic deposition, hydrogenic deposition, wind transport into an area, and beach nourishment. Significant losses are caused by wind transport out of an area, offshore transport, solution, abrasion, dredging, and mining. Human modifications along the coast, including engineering structures, dredging, and beach nourishment, can profoundly influence the patterns and amounts of losses and gains. Human modifications of river basins can also change the amounts and patterns of the supply of incoming materials.

An analysis of sediment supply may involve the investigation of input at a relatively great distance from a given project site. The relationships of supply and transport are complex, in that littoral sediments supplied by one factor, such as river sources, may be transported or removed by another, such as longshore processes. A multitude of techniques including field measurements, charts, maps, photographs and documents, numerical analyses, and computer simulations can be employed. Sometimes, analyses can be facilitated by breaking up the coast into a series of compartments, or cells, to assist in identifying sources and sinks (Carter 1988). The time scale over which data are collected is important, as sediment budgets also reflect the geomorphic variability of coasts, and may be influenced by cyclic and noncyclic changes and long-term trends.

Engineering structures often create obstruction to alongshore sediment transport. Structures most commonly affecting alongshore sediment movement are groins and jetties (Figure 11). Groins are constructed for the purpose of obstructing or retarding alongshore sediment transport to mitigate the effects of erosion. However, they usually cause accelerated erosion of downdrift beaches by cutting off or reducing the amount of material that reaches them (Figure 12). If spaced incorrectly, groins can also cause localized erosion. Offshore breakwaters are being used in some places to reduce the effects of wave erosion, while at the same time allowing alongshore transport processes to continue.

Other engineering structures may create an obstruction to onshore-offshore sediment transport. Seawalls and bulkheads are often built to protect cliffs and dunes from being eroded by direct storm wave activity or by slumping (Figure 13). Where cliffs are composed of unconsolidated material they are an important source of sediment both to the adjacent shore and to the

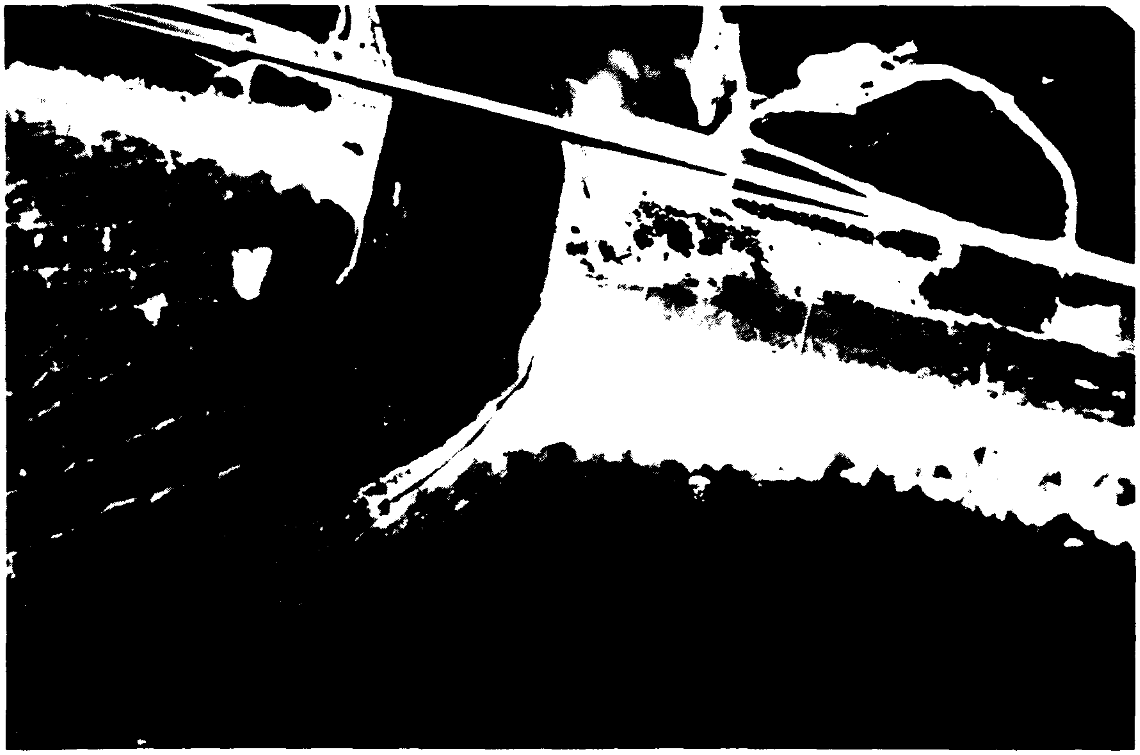


Figure 11. Inlet through a barrier stabilized by jetties. Note the large amount of sand trapped by the larger jetty and recession of the shore across the inlet

Figure 12. Beach protection structures, Cartagena, Columbia (1981). Although the city is built on the barrier beaches which protected the original Spanish anchorage, the present shoreline suffers from a lack of sand, and the beaches are narrow or are entirely missing in some areas

Figure 13. Beach backed by seawall near Galveston, TX. Note riprap added for further protection

alongshore transport system. Therefore, protecting the cliffs can cause net erosion of the adjacent and downdrift beaches.

Structures and activities in river basins also affect the availability of sediment supply to the coast. Sediment supply can be greatly reduced by the building of dams and reservoirs in watersheds because sands, silts, and clays are impounded behind the dams. Therefore, when the rivers ultimately reach the coast, they transport lower sediment loads than they did before the dams were built. As an example, recent erosion of the Nile Delta is attributed to the construction of the Aswan High Dam (Carter 1988). Other structures, such as revetments which reduce bank caving, and levees, which prevent rivers from overtopping their banks, also reduce sediment supply to coastal and wetland areas. Stream diversion, whether a natural event or by design, can cut off important sources of sediment to the coastal areas formerly receiving sediment from those sources. In some circumstances, diversion of streams may supply sediment to an area that formerly had been bypassed. Land-use changes in watersheds, including deforestation, agriculture, and urbanization, may affect the fluxes and timing of sediment supply to coasts.

3 Variable Coastal Features

Coastal features can be examined at a variety of scales. As features are examined in progressively smaller area and greater detail, the morphologic characteristics generally reveal more rapid changes and greater complexity. Some major coastal environments are described herein, with large-scale morphological features being given most attention. These are examined from a variety of perspectives including profile, plan-view, and three-dimensional. Morphological features of smaller scale, such as bedforms, are described in sections where they are often found, including the beach and nearshore zone and tidal inlets.

Various types of geomorphic changes take place in varying environments of the coastal zone, depending upon materials and which processes are locally more important. Large-scale coastal forms generally encompass a wide variety of environments, each exhibiting distinctive processes and responses. For example, the sediments near a barrier island may represent shelf, shoreface, beach/foreshore, dune, back barrier flat, and lagoon environments, each morphologically distinctive (Figure 14). Strandplain coasts and tidal flats contain some, but not all, of the same environments (Figure 14). Coasts strongly influenced by terrestrial fresh water and/or sediment input, e.g., near estuaries and deltas, also have a unique suite of processes, environments, and characteristics.

Material type also affects coastal form. In locations where unconsolidated sediments are available, beaches may form if materials are predominantly sand-sized or coarser, although the processes and characteristics of sand-sized beaches and coarse clastics are distinctive. If abundant sand supply is available, coastal dunes may form, If there are significant storms, overwashing, overtopping, and associated features may occur. Mudflats, marshes, and mangrove swamps may form if materials are predominantly fine-grained. Lithified materials coasts also show distinctive processes, with large-scale morphologies typically including cliffs and shore platforms. Locations where organic deposition occurs typically develop hard, shallow reefs, with forms differing from those found in clastic sediment environments.

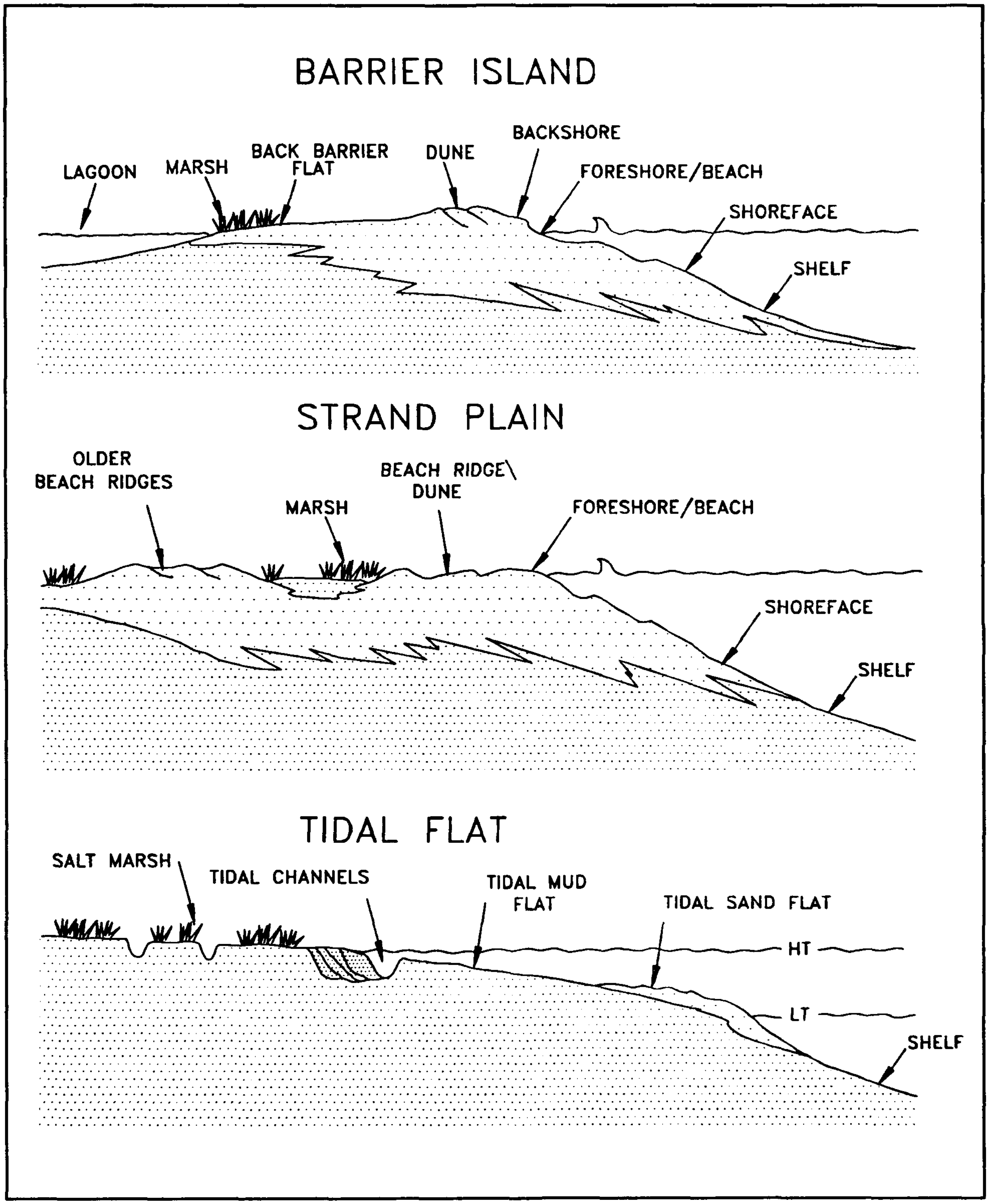


Figure 14. Surface and subsurface environments and variations of barrier islands, strand plain coasts, and tidal flats

Beach and Nearshore Zone

Beaches are accumulations of unconsolidated sediment extending shoreward from the mean low tide line to the inland limit of the littoral zone, where vegetation or a change in relief begins to develop (i.e., coastal sand dunes, beach ridges, terraces, or a cliff line) (Komar 1976). Beaches are among the most variable coastal geomorphic features and the most widely distributed of any of the coastal sedimentary environments (Dolan et al. 1972). Extensive beach development occurs on low-lying coasts where great quantities of sediment are available, primarily at barrier coasts, with the remainder occurring in pocket beaches, lakes, and rock headlands (Davis 1985). Beach sediments generally range from fine sand to cobbles. Finer materials are present on few ocean beaches, because waves create turbulence which keeps fine materials in suspension.

Beaches may hug the coastline or show a number of large-scale detached forms. Barrier islands, which are totally disconnected from the mainland shoreline, are typically fronted by beaches (Nummedal 1983), and separated from the mainland by lagoons, wetlands, or tidal marshes. Spits are quasi-linear subaerial landforms caused by longshore deposition in which beaches separate from the main coastline and project into the deeper waters of estuary mouths or bays. If a spit extends across a bay, it is known as a baymouth barrier. Cuspate forelands are seaward-projecting accumulations of materials. Wave refraction around an offshore island may lead to the development of a tombolo, a beach connecting the mainland with the island.

The beach surface can be divided into two major zones, the backshore, which extends inland from the normal high-tide level, and the foreshore, which is equated with the intertidal zone (Figure 2). The beach and the nearshore zone, which extends from the low-tide level to the seaward limit of bar-and-trough topography, are closely related but show markedly different forms and processes.

Generally, most of the backshore consists of one or more berms or beach ridges, which are flat to gently landward-sloping accumulations of wave-deposited sediments. Gravel and cobble beaches may show one or more storm ridges in the backshore. According to Carter (1988), a beach ridge is a berm that has survived erosion, and there is no real morphological or sedimentological distinction between them. In this study, two types of beach ridges were distinguished, one which was controlled by runup, and one which was the result of ridge stranding. The seaward limit of the backshore is marked by the berm crest, at which the slope changes along a more steeply inclining beach face. Eroding beaches may show a continuous upper foreshore-to-backshore slope and a slightly concave-upward profile.

Morphologic features of the foreshore are more variable and numerous than in the backshore. For example, the beach face slope may be inclined

from 1 deg to 30 deg, depending upon the sediment character of the foreshore and the processes acting upon it. Steeper beach faces generally occur with coarse materials and in higher energy environments. A small and sometimes subtle step, typically marked by a concentration of shell debris and/or coarse sediments, may be present at the line where the waves plunge before surging up the beach face. A sandbar and landward trough, known as a ridge-and-runnel system, may also be present at times (Figure 2). The ridge-and-runnel system is most common on beaches with an abundance of sand. Seaward of the ridge near the low-water line, there may be a nearly horizontal morphologic zone called a low tide terrace.

Rhythmic topography changes also occur in varying scales on the foreshore. Beach cusps (spacing of 10-30 m) and giant cusps (spacing of 100-200 m) are examples. Both cusp features have a similar morphology and a non-tidal genesis, with beach cusps being related to accretionary processes during swell conditions, and giant cusps being formed during storms under erosional wave conditions. Spacing of beach cusps may be related to rip current spacing, the wavelength of edge waves, and the spacing between waves arriving at the beach from different directions. Larger rhythmic features called beach protuberances (100 m-100 km) are characterized by subtidal components and sandbar movements. Non-rhythmic morphologic features, forming aperiodic protuberances, may also occur at the shoreline, often in response to the longshore drift patterns associated with cells. Other non-rhythmic features, such as salients, may form in sheltered areas behind rock outcrops or offshore breakwaters.

Morphologic features of the nearshore zone include a number and variety of stable and ephemeral subtidal bar forms. These often show relationships to beach topography. The formation of submarine bars is favored by, but not entirely dependent on, a gentle shoreface slope, low tidal range, ample sediment supply, and a low incidence of long swell waves. Bars are absent only where these factors are lacking, and are less stable in shallow water than in deeper water.

Several types of submarine bars have been recognized. They are differentiated mainly on the basis of plan view shape, alignment, and continuity. Longshore bars are the most common, consisting of a 1- to 4-meter relief linear ridges aligned parallel to shore. Their continuity is broken only by narrow rip channels which may migrate alongshore. The origin of longshore bars is believed to be related to breaking waves, and most bars seem to be located at the breaker line (Miller 1976). Where multiple longshore bars exist, their positions are often associated with the breaker lines at high- and low-tide stages, multiple breaker lines, or breaker positions for various wave characteristics.

In the nearshore zone, crescentic bars are the next most common form. They are shaped convex-seaward, and may be in-phase or out-of-phase with shoreline protuberances. Lunate bars form a half crescent, with the bar initially extending from the shore protuberance, then bending to a more or

less shore-parallel orientation pointing in the direction of the longshore current. Transverse bars are oriented perpendicular to shore and are attached at beach protuberances.

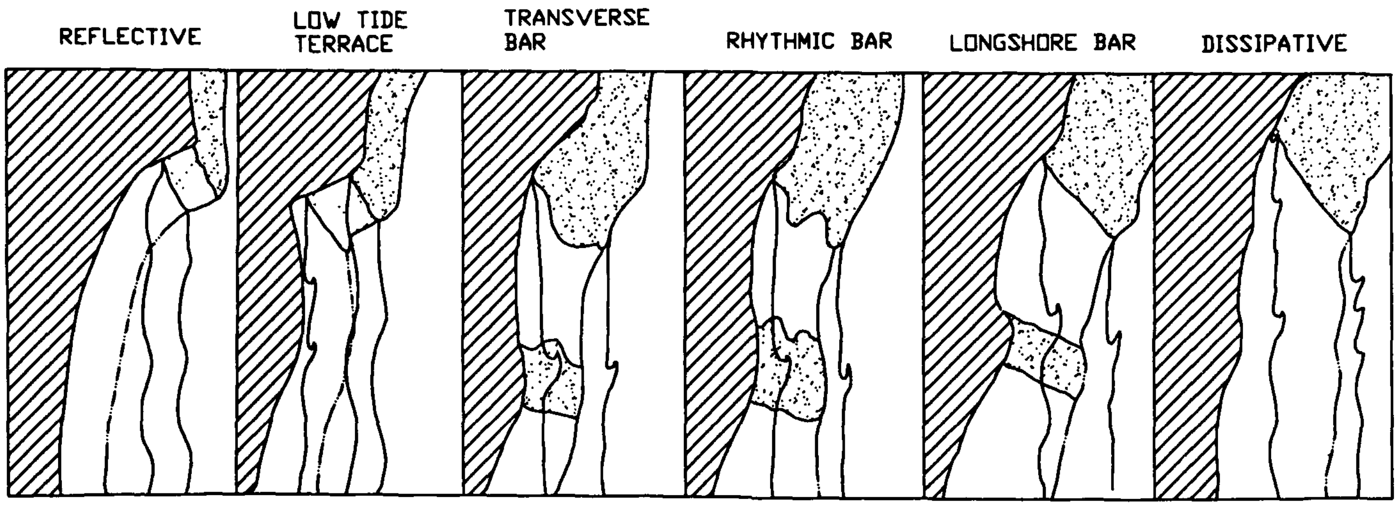
Process-response mechanisms in the beach and nearshore environment have been integrated into a number of distinct morphological states, ranging from dissipative to reflective with several transitional morphodynamic regimes (Wright et al. 1979; Short 1979; Wright and Short 1983). The system interrelates waves, currents, morphology, sediment size and sorting, and sediment transport (Figure 15). Dissipative profiles are characterized by low gradients and wide surf zones, multiple parallel bars, and suspended load transport. Reflective systems have steep beach faces with surging breakers, edge waves and widespread cusps, bed load transport, and an absence of nearshore bars and rip currents. Intermediate regimes incorporate elements of both domains, progressing through the transitional intermediate states from longshore bar and trough to rhythmic bar and trough, to traverse bar and rip, to ridge and runnel.

On predominantly sandy coastlines, beaches and the nearshore region continually respond to ever-changing winds, waves, tides, and currents by showing adjustments in profile and morphologic features according to beach type and environmental conditions. The day-to-day changes can be notable. On the beach and nearshore, when wave energy is low or moderate, there is overall net onshore transport and constructional activity.

As energy increases, long-period, steep-profile storm waves may produce considerable erosion of the beach and nearshore environment. Longshore bars, for example, can shift position quickly, moving offshore at rates of 30 m daily. In response to less steep waves, bars move more slowly (Birkemeier 1985; 1987). Many types of morphologic changes respond to tidal cycles. One example is the migration of erosion and accumulation zones on a diurnal or semidiurnal basis with tidally driven water table changes (Duncan 1964).

In addition to the day-to-day fluctuations in beach morphology, longer term cyclic and unidirectional effects occur on sandy beaches. The principal cyclic effects are usually related to seasonal variations in dynamic factors, which in turn create distinct cyclic changes in beach morphology and sediment characteristics. Thus, survey data should ideally cover different seasons in order to indicate the range of values that may be encountered during a year.

During periods or seasons when frequent or severe storms occur, (typically winter), sand is eroded from the beach, causing the profile to become lower and narrower. Offshore bars may develop or enlarge due to the addition of material from the beach. With the return of fair weather conditions, the beach tends to recover and all or a part of the material in offshore bars may return to the beach. Occasionally, overtopping and overwashing occur on sandy coastal barriers, when water and sediment may pass over the barrier crest and settle on the landward side. Overwashing events are generally noncyclic, preventing the beach from recovering to its initial condition. It is rare that



# OF WAVES IN SURF ZONE	1	2	3
SURF ZONE WIDTH	10m	10-100m	> 100m
BREAKER TYPE	PLUNGING/COLLAPSING	SPILLING/PLUNGING	SPILLING
∢ OF WAVE APPROACH	STRONGLY OBLIQUE (10-90°)	SLIGHTLY OBLIQUE (0-10°)	SHORE-NORMAL
NEARSHORE CURRENTS	LONGSHORE	OBLIQUE OR MEANDERING CELL	CELL CIRCULATION
BEACH SLOPE	STEEP (<3°)	INTERMEDIATE (1-3°)	GENTLE (<1°)
GRANE SIZE	BOULDERS,GRAVEL, COARSE SAND	MEDIUM SAND	FINE SAND,SILT
SEDIMENT TRANSPORT MODE	BED LOAD	SALTATION,MIXED LOAD	SUSPENDED LOAD
SEDIMENT TRANSPORT DIRECTION	ALONGSHORE	MIXED	ONSHORE/OFFSHORE

SELECTED MORPHODYNAMIC ATTRIBUTES

Figure 15. Three-dimensional morphodynamic classification of beaches and selected associated attributes (after Wright and Short 1983)

material moves from the back to the front of a barrier. However, seaward-flowing currents may cross barriers if landward water levels become elevated, a phenomenon sometimes termed the storm surge ebb-residual flow (Hayes 1967).

Coarse-grained clastic beaches are distinctive in terms of their grain size, wave energetics response, and profile characteristics, but are also distinctive in sediment transport and coastal evolution. Such beaches are usually morphologically reflective, dominated by plunging breakers with no surf zone, strong longshore bedload transport in the swash zone, and seepage as a major process (Carter 1988). While sand beaches undergo tidal and seasonal changes, gravel beaches have arrested swell profiles, which are maintained through most storm periods but can show aseasonal responses. Coarse clastic beaches may show alongshore grading. Sediment transport processes are poorly understood in these cases, but are likely related to the wave parameters as on sand beaches.

Longer term cyclic and noncyclic events affecting beach and nearshore morphology may extend over a period of decades or centuries. Variations in climatic patterns, wave characteristics, sediment supply, and relative sea level, as well as the lasting effects of coastal engineering works, inlets, and extreme events may cause long-term cyclic and noncyclic changes. Studies of historical maps, charts, and aerial photographs are valuable to show relatively long-term trends of such factors as shoreline or shoal migration, and bathymetric changes that should be considered in project design.

If a beach has been subject to environmental forces for an adequate period, the beach profile will respond to both the long-term and short-term changes in a manner that tends to restore an equilibrium profile. In equilibrium, the amount of sediment deposited by waves and currents will be balanced by the amount removed by them. Some researchers have proposed that the slope of natural profiles fits the following equation: $h(x) = Ax^m$, where A is a scale factor primarily dependent upon sediment characteristics and m is a shape factor, proposed to be 2/3 (Bruun 1954; Dean 1977; Dean and Maurmeyer 1983). While the equilibrium profile may be disturbed by unusual and exceptional conditions, such response models can assist in predicting future shoreline positions and in interpreting shoreline history.

Over historic, and particularly over longer time scales, shoreline response to sea level rise may follow one of three generalized models (Figure 16). The erosional response, or Brunn Rule model, assumes offshore dispersal of eroding shoreline materials such that the rates of sea level rise and sea bed rise are commensurate. In the rollover model, a transgressive barrier moves landward at a rate controlled by the rate of sea level rise (Dillon 1970). The barrier overstepping model suggests that the barrier may be drowned, remaining on the shoreface, as sea level rises above it.

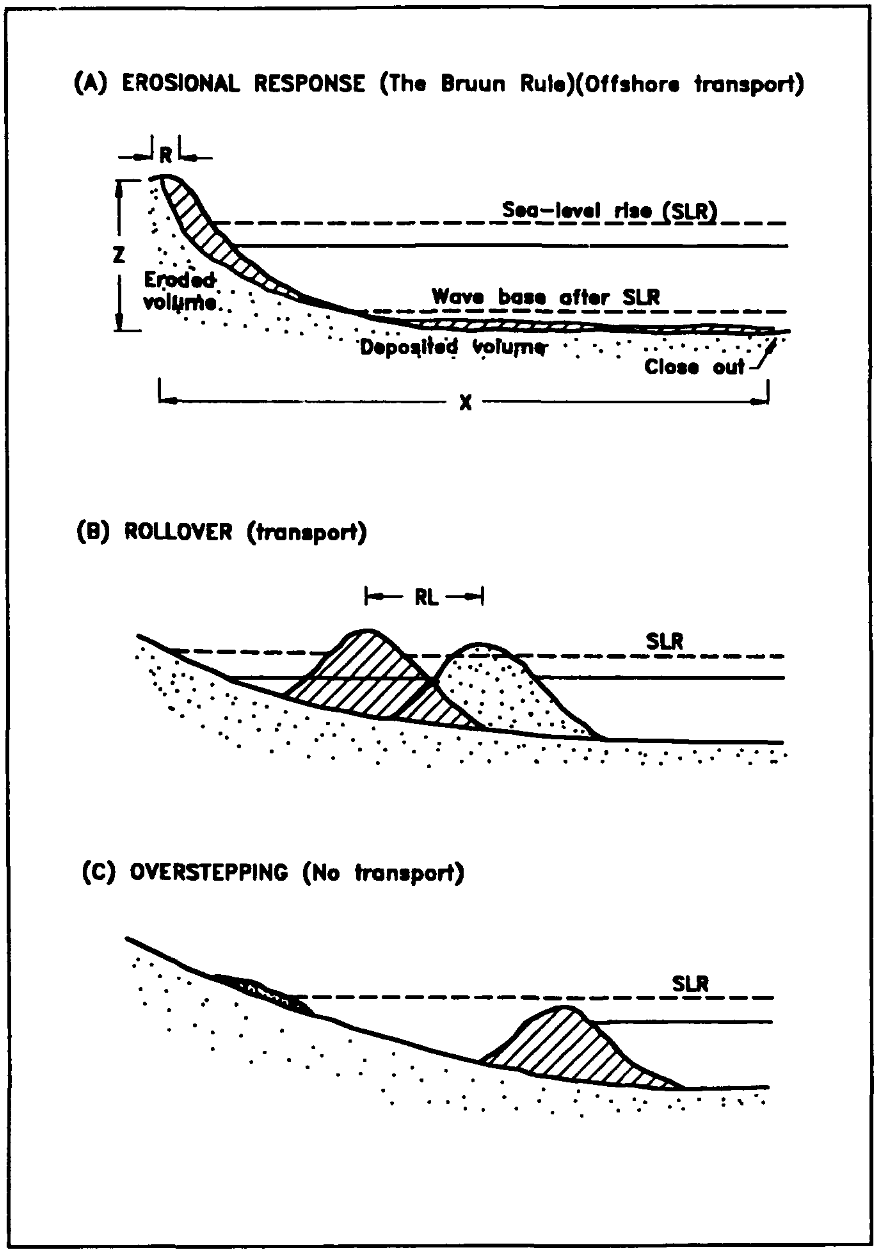


Figure 16. Three models of shoreline response to sea level rise: (a) Erosional response model/Brunn rule assumes offshore dispersal of shoreline materials; (b) Island rollover model assumes barrier migrates landward according to rate of sea level rise; (c) Barrier overstepping model assumes submergence in place with barrier remaining on shoreface (after Carter 1988)

Lithified Coasts: Cliffs and Platforms

Rocks and sediments that are semi-consolidated or consolidated may form vertical or steep cliffs that constitute a marked break in slope between the hinterland and the shore. If marine processes are given adequate time, the cliff may be fronted by a gently sloping shore or intertidal platform where debris may rest (Figure 17). The slope and recession rates of the cliff faces depend on a number of terrestrial and marine processes, as well as the geotechnical properties of materials comprising them, including grain size and degrees of consolidation. Recent volcanic or loosely consolidated Quaternary sediments show the greatest erosion rates (Sunamura 1983).

Marine processes remove cliff materials directly by attack at the cliff base, and indirectly by undercutting, causing failure or mass wasting of the overlying rock. The debris produced by these processes may then be transported by a variety of marine processes. Resulting accumulation, erosion, or cliff modification depends on the relationship between the relative rates of supply and removal at the shoreline (Pethick 1984).

If marine processes are capable of removing slope debris much faster than the rate of debris supply, then, in most cases (depending upon the structure and lithology of the rock) the slope will retreat parallel to itself. When the supply of debris far exceeds the capacity for removal, basal debris will accumulate at the angle of repose (the maximum angle of slope at which loose cohesionless material rests). Between these extremes, a variety of slopes ranging from debris angle of repose to vertical faces are possible.

Typically, profiles of steeply sloping shorelines are highly reflective, whether the materials are consolidated or unconsolidated. Currents are shore-parallel and unidirectional, and bars are generally absent. Longshore sediment transport rates are high, and most materials are moved away by currents. Thus, along cliff coasts, the rates of supply must generally exceed those of lower slope coasts to produce beaches.

As with beaches, lithified coasts show irregular shorelines. Headlands and bays may be related to the submergence of a hilly or mountainous topography by a rise in sea level. Differential erosion and weathering may also lead to the development of alternating headlands and bays on rocky cliff coasts. Once formed, the presence of prominent headlands on rocky coasts influences waves and tides, sediment dispersal and deposition, and shoreline evolution. Headlands influence refraction, causing wave ray convergence. Recent research has shown that headlands may protrude into tidal flows, causing tidal eddies, and in some cases providing a stagnation zone where offshore shoals can form. These, in turn, can alter the wave energy environment, creating nonuniform wave attack, and altering the spatial patterns of cliff erosion (Carter 1988).

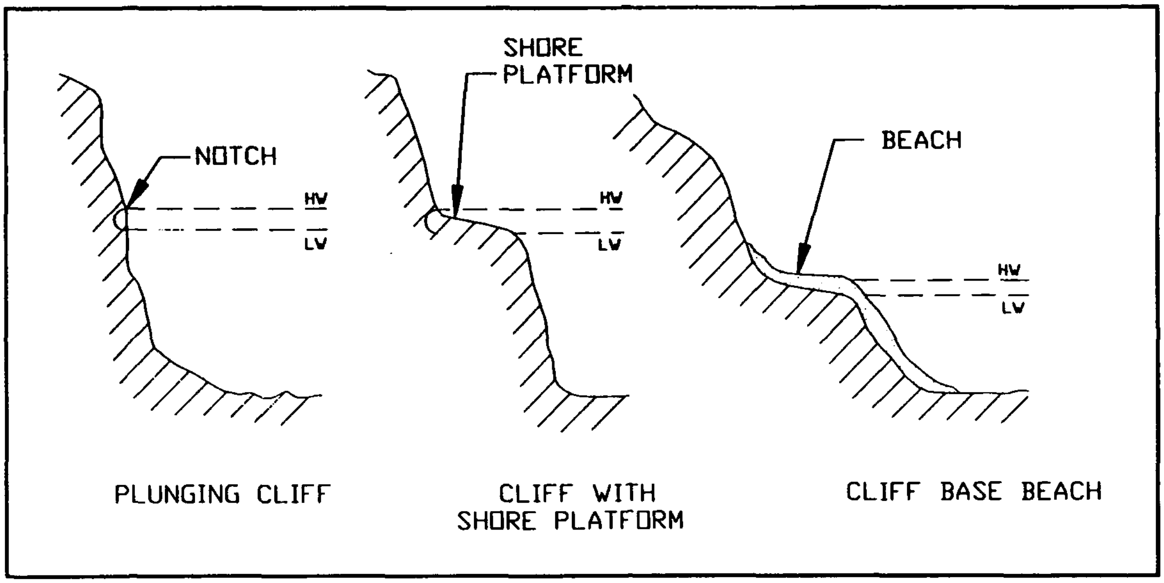


Figure 17. A variety of cliff morphologies in cross section. Plunging cliff is steeply sloping and possibly shows a notch. Cliff with shore platform may develop from increased notch development and mass movement of overlying material. Cliff base beach may develop from cliff with shore platform if sediment supply exceeds transport of materials

Coasts of consolidated materials resist wave attack because of compressive, tensile, and cohesive properties. Thus, because they are primarily erosional rather than depositional in nature, they show distinctive morphologic features in comparison with unconsolidated coastal sediments. Sustained wave attack may etch notches in rock, the deepest notches occurring where dynamic pressures are greatest. Multiple notches may form if there are various dynamic wave horizons, most likely associated with changing water levels. Notches may become sea caves along zones of structural weakness. Variations in resistance may also lead to the development of sea stacks and sea arches seaward of the cliffs.

Coastal slope recession typically leaves behind a shore platform, a surface that is quasi-horizontal or of low slope angle, marking the lowest level to which the erosion reached. Maximum width of such platforms is about 1 km (Flemming 1965), and their profiles may be linear, concave, or convex. A variety of processes act on platform surfaces, including abrasion, mechanical wave erosion, weathering, and solution, which vary depending upon the nature of the processes, rocks, structure, tides, age, and history.

Abrasion from sand and shingle is concentrated on the upper shoreward section of the platform since these abrasive materials are generally absent on the lower levels (Robinson 1977; Trenhaile 1980). Considered by some to be the most important, mechanical wave erosion may assist in breakup and platform surface lowering by wave shock, wave hammer, air compression, and other forms of dynamic pressure release (Trenhaile 1980). Several weathering processes may also be involved, particularly those associated with alternate

wetting and drying of the platform surface. This water-layer weathering includes physical processes such as expansion, swelling, and salt crystallization, and chemical processes such as hydration and oxidation. Wave erosion may proceed more rapidly in weathered material, with subaerial processes at the water table level possibly coming into play. In tropical calcareous areas, solution of rocks and bioerosion appear to be important processes in shore platform development.

While the complexities of platform morphology are not well understood, the shore platform functions as a dissipator of wave energy and as a pathway for sediment transport at the cliff edge. The platform slope may thus influence the cliff, in that steep slopes would create increased longshore sediment transport and promote cliff recession, whereas gentle slopes would not allow sufficient longshore transport and beach progradation would result (Bradley and Griggs 1976). The critical platform slope would be dependent upon the shear stress at the bed by wave forces and the strength of materials making up the shore platform.

Organic Reefs

Reefs, which are formed mainly from biogenically produced carbonates, are important structures in tropical waters. Like lithified coasts, they form hard bottoms, but several unique forms develop in such environments. Several types of marine organisms are capable of precipitating calcium and other carbonates in skeletal and non-skeletal forms, but reefs of coral skeleton are the most common. Corals grow most successfully in shallow, warm, mud-free waters of moderate salinity, characteristically between latitudes of 30° N and 30° S where such conditions occur.

Four major forms of large-scale reefs have been identified: (a) fringing reefs, (b) barrier reefs, (c) atolls, and (d) table reefs (Stoddart 1969). Fringing reefs are connected with the land, whereas barrier reefs are separated from land areas by a lagoon which may be several kilometers wide. Minor reef shapes include ring-shaped forms on banks and shallows, and reef knolls or patch reefs growing in lagoons. Like barrier reefs, atolls rise from deep water, enclosing a lagoon, but unlike barrier reefs, they do not enclose land within them. Table reefs rise from the sea floor as a shallow bank, being capped with reef growths that appear different from the other forms.

The reef surface has a high roughness, and will change local morphodynamics by affecting the energy of incident waves and by increasing turbulence. Reef form and structure are considered to be linked largely to wave climate, and may show process gradients. Studies on the Grand Cayman reefs show that the outer reef is dominated by wind-driven currents, the inner by high-frequency waves, and the lagoon by deepwater waves and tides (Roberts, Murray, and Suhayda 1975).

Coastal Dunes

Coastal dunes (Figure 18) occur where there is a sufficient sediment source, onshore winds of sufficient velocity to transport the available material, and a flat or low-relief space inland of the coastline to accommodate dune formation. Sources of dune material are usually the dry portion of sandy beaches. The wider the beach, the greater the surface area exposed to wind transport and, consequently, the greater the volume of material available to form dunes. Flat, dissipative shorelines, where large volumes of sand are stored subaerially at low water, are highly suited for foredune development (Short and Hesp 1982).

Figure 18. Partly vegetated coastal sand dunes. Eastern Alabama near Florida/Alabama state line (March 1991). This area was devastated by Hurricane Frederic in 1979

Tide range and type can be of considerable importance in beach-dune interactions. On beaches where tide ranges are high, more sand is uncovered at low water, effectively increasing the size of the source area. Tide type is also important because the wind velocity necessary to move sediment is significantly increased when the sediment is moist or wet. Because of this, more sediment is available in areas where diurnal tides occur, since a longer period of time is available for the foreshore sand to dry out between high tide stages.

Effective transportation of sediment by wind occurs when the wind reaches a threshold velocity that is determined not only by sediment characteristics, but by moisture content, slope, radiation and energy balance factors, chemical precipitates, and vegetation (Carter 1988). The threshold velocity is significantly increased for moist material. Also, the seaward slope of the beach surface increases threshold values in accordance with the steepness, because the particles must move upslope to reach the dune area. Sand transport from the beach to dune areas occurs only when wind blows in an onshore direction, whereas sand transport from major inland sources occurs only when wind blows in an offshore direction. Thus, the frequency of onshore and offshore winds, their relation to local sources, and their velocity, are important factors in dune development.

Dune formation requires an open space inland of the coastline that can be reached by windblown sand from the beach. A relatively high cliff or steep slope at the coastline can prevent dune formation by blocking the transport path. In such cases, windblown material may accumulate at the base of the cliff or slope, but this material will be periodically eroded during high water levels, and it will develop into true dunes.

The stability of dunes varies greatly and depends primarily on the amount of vegetation cover. Dunes found in arid climates are usually not vegetated and tend to be mobile. Coastal dunes in non-arid climates are likely to support some vegetation cover and be more stable. However, unvegetated and thinly vegetated dunes can occur in any climate. Colonization of dunes by vegetation not only depends on whether an area has sufficient moisture, but whether the grasses and other plants are salt-tolerant and can respond to conditions of rapid sand accumulation. The types of dune grasses found in an area show spatial variety according to the types of species in an area, and rapid temporal successions are often present (Carter 1988; Chapman 1964; Goldsmith 1985; Woodhouse 1978).

Vegetation cover is a factor in the stability of dunes because the plants can both decrease the mobility of sediments and increase the likelihood of deposition by changing local aerodynamic conditions. Many dune plants have long roots, rhizomes, and runners that help hold sand in place (Figure 19). The presence of dense vegetation, in turn, at normal wind speeds can displace the aerodynamic boundary of the wind velocity profile upwards. This process then provides a net downward momentum flux, which promotes sediment trapping.

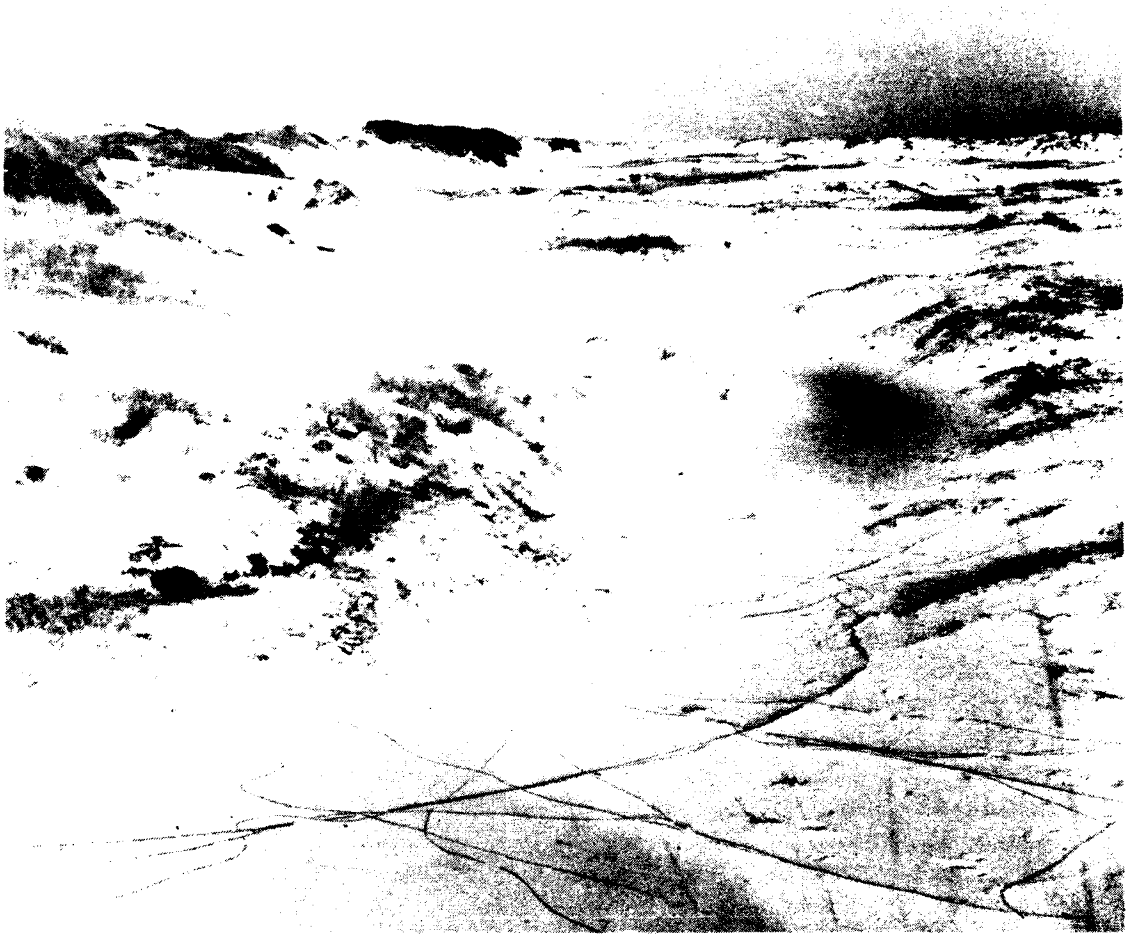


Figure 19. Rhizomes help hold sand in place and spread the colonization of dune grasses. Eastern Alabama near Florida/Alabama state line (March 1991)

Because dunes are effective in sediment stabilization and trapping, dune vegetative cover and form are interrelated and can be classified as such (Short and Hesp 1982). Densely vegetated dunes are associated with fixed, shore-parallel dune ridges. As vegetation decreases, dune patterns range from discontinuous ridges to hummock dunes to small blowouts. Minimally vegetated dunes show residual knolls, barchanoid forms with crescent shapes and downwind horns, and transverse blowouts or hollows.

Dune vegetation can be damaged or destroyed by various diseases and natural and human-induced disturbances. Climatic and meteorologic stresses, particularly droughts and storms, fire, grazing by wildlife and domesticated

species, and foot and vehicular traffic can alter the morphologic and ecologic conditions of dunes.

Because dunes help protect inland areas from storm damage, many communities promote dune stabilization and protection. For dune construction, beach nourishment is often practiced in conjunction with measures to stabilize and trap sediments, (e.g., fencing, vegetative planting, or the placing of other obstacles). These measures, as well as the construction of walkways to reduce trampling, are also used for dune stabilization and protection.

Lithified or fossil dunes occur in some places where there are calcium carbonate particles in the dune material. Climatic factors promote leaching and reprecipitation of calcium carbonate, which may allow dunes to lithify under favorable conditions. Modern coastal dunes that have become lithified occur largely in tropical climates where there is a high level of calcium carbonate in the coastal sediments and alternating wet and dry periods.

Back Barrier and Lagoons

During storms and high water levels, the sea may breach a dune ridge, bringing sediment from the coastal area further inland. Some scientists call the phenomenon overwash, and the product is called washover (Carter 1988). Overwashing may result in a distinctive set of breach-throat-fan landforms, forming a washover channel through the dune and a washover fan. This washover fan develops on the landward side of a barrier and spreads over parts of the backshore, back barrier flat, back barrier marsh, or shallow lagoon (Figure 20). Storm frequency, overwash volume, dune susceptibility, and barrier height are important factors in the development of washovers. Low-lying or low-profile barriers in areas of frequent, severe storms have numerous washovers. Storm surge ebb-residual flow, which occurs more infrequently than overwash, may result in seaward currents and landforms.

Washover channels are distinguished from inlets in that their elevation is above mean sea level. The patterns of breaches may be related to a combination of marine or non-marine processes. Washover deposits provide an important mechanism by which marine transgression may take place, not just for sand-dominated, but also for gravel-dominated barriers (Carter and Orford 1984). Overwash features are particularly common in the Gulf of Mexico coast and eastern seaboard of the United States. These areas experience severe extratropical and tropical cyclones with some frequency.

Coastal lagoons are shallow water bodies, often running parallel to the coast and connecting to the open sea with an inlet. Lagoons differ from tidal flats in that they remain water-filled even at low tide and are separated from the open sea by sand bars, barrier islands, or reefs. Lagoons have been

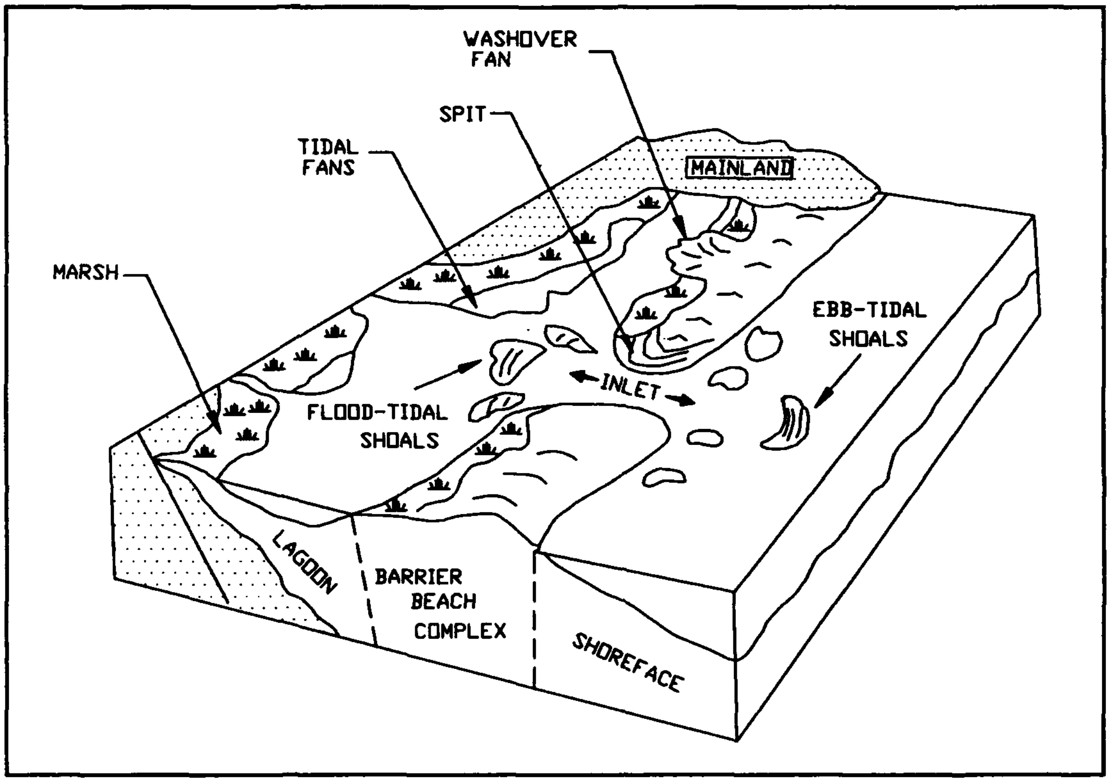


Figure 20. A three-dimensional view of some features commonly associated with a barrier island system, including the back barrier, overwash fans, and lagoons

classified according to hydraulic balance, where inflow is equal to, greater than, or less than outflow. The volume of water being exchanged is known as the tidal prism (Carter 1988). The salinities of lagoons are highly variable, depending upon water exchange with sea water and the amount of inflowing fresh water. Sediment brought into lagoons may be washed over from the barrier island, blown into the lagoon from the barrier island, introduced by tidal currents through inlets, or transported from the mainland areas by rivers. Generally, the coarsest sediments are found closest to the barrier.

Mudflats, Salt Marshes, and Mangrove Coasts

Mudflats, despite their name, are not entirely flat, nor do they consist entirely of mud. They occur predominantly in areas of medium to large tidal range, areas sheltered from the effects of wind-driven waves, or in areas of abundant suspended sediment supply (Pethick 1984), including in the vicinity of deltas. Mudflats show a marked break in slope about mid-tide level, below which the surface slopes steeply toward the low-water mark. Moving seaward and downslope, sediment size increases, with upper mudflats being replaced by sandy mudflats at slightly lower levels. These sandy mudflats in turn change into more steeply sloping sand flats below mid-tide level. Fine-

grained sediments may be supplied from marine, coastal cliff, fluvial, or estuarine sources.

Mudflat morphology is related to tidal and sediment processes, and the break in slope at mid-tide generally reflects the position of the maximum tidal velocity (Pethick 1984). The upper part of the mudflat surface progressively increases in height due to the accretion of sediments, so that the period of inundation of each tide is successively reduced. Eventually, the mudflat may become exposed long enough to allow vegetation to colonize, leading to the development of salt marshes.

Salt marsh development usually begins with the deposition of mud on a sand surface followed by the establishment of algae salt-tolerant plants like *Spartina Patens*, *Spartina Alternifora*, and eelgrass near the high-water mark. The vegetation and seaweed may assist in trapping increasing amounts of sediment. Increased vegetative sediment trapping leads to the upward and outward building of hummocks, and to reduced wave heights and energies. Rates of marsh sedimentation are controlled by sediment availability and vegetative trapping, and also by the magnitude and frequency of various factors affecting water level elevation, including waves, tides, and surges. With increasing elevation, however, there will be decreased frequency of inundation, and thus reduced rates of upward accretion during storms.

Salt marshes are well-developed during periods of relative sea level stability, and typically occur on marine delta plains, behind barrier beaches, in depressions, embayments, and other irregularities of the coast. Salt marshes trap muddy sediments in low-energy tidal situations along protected sections of extratropical coasts, comprising 1- to 2- km bands along the Atlantic and gulf coasts of the United States. Numerous organisms are uniquely adapted to such conditions, and the salt marsh substrate records contain many details that are of significance in paleoecology and environmental reconstruction.

Mangrove swamps occupy settings similar to salt marshes, except that they occur in lower latitudes, between 30 deg N and 30 deg S. Mangroves include several species of low trees and shrubs, and are characterized by an entanglement of arching prop roots that facilitates trapping of fine sediment. Typically, they are composed of higher amounts of organic debris than marshes. Mangrove growth is favored by tidal submergence and high tidal range, low coastal relief, saline or brackish water, abundant fine sediment supply, and low wave energy. The most notable area of mangrove development in the United States is in southwest Florida.

Estuaries

Estuary definitions vary, but all estuaries share several important attributes. They are semi-enclosed water bodies in which tidal exchange and fresh water

from land drainage combine, typically resulting in hydrodynamic, turbidity, and salinity gradients. In contrast with deltas, estuaries occur at the mouths of rivers that have low sediment loads in comparison with dissipative forces (Nichols and Biggs 1985). Lagoons behind barrier islands are also classified as estuaries.

Estuaries may develop under a variety of climatic and topographic conditions, including valleys and coastal embayments that have been submerged. Many estuaries were formed mainly during the most recent deglaciation and associated sea level rise. These include most of the coastal plain estuaries, some of which have been closed by barrier beaches. Fjords are a special class of estuaries created by the scouring action of glaciers. Fault-block estuaries may be controlled by local or regional structure.

Estuaries function mainly as sinks or traps, although those that are largely filled may be sources of sediment. Their long-term survival depends on changes in the volumetric capacity for storage (due to eustatic and tectonic factors) and on sedimentation rates from inland and coastal sources. Estuarine sediments are derived from a number of sources including the watershed, the Continental Shelf, local erosion, biological activity, and the atmosphere. Sedimentary processes in estuaries are controlled by tides, waves, and meteorological forces, as well as river inflow. Within an estuary, the processes may be dominated by estuarine-fluvial, estuarine, or estuarine-marine activity. As in other parts of the coast, estuaries show temporal variations with cyclic and noncyclic processes, and spatial variations in process, material, and form.

Estuaries have been classified according to the mixing processes caused by the density differences between fresh- and saltwater masses (Pritchard 1967). Salt wedge estuaries are highly stratified, with fresh water from river discharge floating over the denser saltwater, a condition manifested by near-horizontal isohalines. The freshwater layer thins oceanward, and vertical advection is the primary mechanism for mixing across the freshwater/saltwater interface. A partially mixed estuary shows increased tidal influence, to the point where river discharge does not dominate circulation. Mixing is caused by both vertical advection and the increased turbulence associated with tidal currents. The differences between surface and bottom salinities are approximately constant over the estuary, and the isohalines are at an angle. A fully mixed estuary is vertically homogeneous because tidal mixing eliminates vertical density stratification.

Shoreface

The shoreface is just seaward of the surf zone, and appears as a concave-upward surface with a slope on the order of 1:200 (Niedoroda, Swift, and Hopkins 1985). The shoreface configuration, particularly the upper part, plays an important role in the modification and transformation of waves

approaching the shore. In addition, data from many areas suggest that under certain circumstances, there is an interchange of sediments between the upper shoreface and the beach. Although the interchange occurs on a seasonal basis, it may be unidirectional at times.

Because of the cost and difficulty of obtaining data and sediment samples from offshore areas, less is known about the shoreface zone than the adjacent shore and coast. However, repetitive profiles indicate that shoreface morphology is variable over the long term (Moody 1964), and that episodic sediment transport, with suspension of sediments at least 1 m above the sea floor, occurs in response to changes in waves and currents (Young et al. 1982; Vincent, Young, and Swift 1983).

Everts (1978) analyzed 49 composite shoreface profiles based on 441 measured profiles from the Atlantic and gulf coasts of the United States. Of the 49 composite profiles, only three, all from northwestern Florida, did not contain a definable shoreface slope. In general, the profiles show the shoreface to have a concave-upward slope and to be significantly steeper than the usually planar shelf floor, perhaps representing a cutoff point of significant active modification of the profile by waves and currents.

Shoreface variability is more pronounced on the upper shoreface because of the decrease in wave and current forces with increasing depth. The most pronounced short-term effects appear to be seasonal, with the more energetic winter regimen tending to move material from the beach to the shoreface and summer fair weather conditions tending to move material onshore. Lower shoreface deposits are disturbed less often and probably become active only during major storms.

On many transgressing barrier coasts, the barriers are overriding their own back-barrier deposits, which may crop out on the beach or the shoreface. Evidence of outcropping back-barrier deposits on transgressive coasts include blocks of salt marsh peat and shells of back-barrier fauna such as *Crassostrea virginica* (Gmelin). Most back-barrier sediment is fine-grained and tends to move rapidly offshore to deeper water where it is more stable. Coarser sand may also occur in back-barrier deposits in the form of storm washover deposits and flood-tidal shoal complexes adjacent to relict inlets. In most cases this material was originally derived from beach sediments moving in the alongshore drift or from relict river channel deposits. Thus, these deposits are more stable and are not likely to increase shoreface recession rates.

Usually, shoreface areas are primarily composed of unconsolidated sands. However, many shoreface zones are underlain by consolidated material such as rock or reefs, glacial till, or clay. These shorefaces may be more stable and may not follow the equilibrium profiles of primarily depositional shorefaces or of those underlain by ancient unconsolidated material. Admixtures of silt and clay are not uncommon, especially on the lower shoreface, where outcrops of fine-grained back-barrier deposits often occur on transgressive coastlines.

Generally, sediment grain size of shoreface deposits decreases in an offshore direction because of wave energy distribution along the profile. Exceptions occur where coarse materials that occur in deeper shoreface areas are left as finer sediments and moved onshore, or where outcrops of relict substrate material are exposed on the sea floor.

Inlets

An inlet is a small, narrow opening, recess, or indentation into a coastline or a lake through which water penetrates into the land (Bates and Jackson 1980). Although inlets range in size from the narrow short breaches in sandy barrier islands to the wide entrances of major estuaries, most geologic studies concern tidal inlets, which are interruptions in barrier beaches maintained by tidal flow (Fisher 1982). Tidal inlets are most characteristic of sand-dominated barriers, and may originate as natural interruptions in a developing barrier beach or baymouth spit, or as breakthroughs caused by storm waves (Fisher 1982). Gravel-dominated barriers tend to lack inlets and tidal passes because they have greater structural stability and more seepage than do sandy barriers (Carter and Orford 1984).

Tidal inlets usually consist of a gorge or throat, and several shoals including a shallow one flanking the gorge, a landward flood-tidal delta shaped mainly by flood-tidal currents, and a seaward ebb-tidal delta shaped mostly by ebb-tidal currents and waves. Sand in the alongshore drift system, which is intercepted by the inlet tidal currents, supplies the inlet shoals (Figure 21). Nevertheless, some material may bypass the inlet. Morphological features associated with tidal deltas include small tidal spits and topographically high semicircular ridges, which extend across the breadth of the sandbank as flood or ebb shields (Hayes 1980).

Inlets exchange water between the ocean and back barrier during each tidal cycle. Therefore, several aspects of inlet morphology are related to tidal processes. Inlet spacing, for example, decreases with increasing tidal range, and inlet migration is influenced by tides and the quantity of littoral drift.

Microtidal barrier inlets tend to be widely spaced and ephemeral, migrating rapidly with the longshore drift direction. The primary process is by erosion of the updrift margin of the inlet and deposition on the downdrift margin (Galloway and Hobday 1983). Deeper mesotidal inlets are less subject to longshore migration, especially if they are incised into harder underlying strata. At some sites, updrift inlet migration has been noted (Aubrey and Spear 1984; Carter 1988).

Variations in tidal delta morphology can also be related to differences in tidal range, and thus the tidal prism, and to differences in wave-energy flux impinging upon an inlet (Boothroyd 1985). Microtidal areas are thought to

Figure 21. Aerial view of unstabilized barrier inlet showing shallow parts of the flood and ebb tidal deltas

have poorly developed ebb-tidal deltas and relatively larger flood-tidal deltas because of the dominance of wave energy over the small tidal prism. Along mesotidal areas, ebb-tidal deltas are more prominent, but are more elongated and extend farther seaward in areas of lower wave energy.

On a smaller scale, variations in bedforms, or deviations from a flat bed, and accompanying sedimentary structures are associated with inlet hydrodynamics including the effects and interaction among tidal currents, wave characteristics, and longshore drift. Bedforms show varying wavelengths and amplitudes, and have been divided into categories largely on the basis of wavelength. Ripples are small bedforms with spacings to 60 cm, which can be generated by both waves and currents.

In the past, terminology for large-scale bedforms has been confusing and it has recently been suggested that large bedforms be called subaqueous dunes (Ashley 1990). First-order descriptors of such subaqueous dunes include size and shape. Descriptors based on spacing include small (0.6-5 m), medium (5-10 m), large (10-100 m), and very large (> 100 m) dunes with corresponding heights. Shape is distinguished by two-dimensional and three-dimensional descriptors. Second-order descriptors should be used where feasible to characterize superposition (simple or compound bedforms) and sediment type (size and sorting). Third-order descriptors characterize bedform morphology, bedform behavior, and flow.

Bedforms are sensitive to flow velocity and somewhat independent of depth, which allows them to serve as a powerful tool in estimating flow velocities in estuaries when field current measurements are not possible (Boothroyd 1985). The shape of bedforms can also vary in response to increasing flow strength (Hayes and Kana 1976). The orientation of these shapes and associated slipfaces also provides clues to flow direction.

Inlet stability plays an important role in coastal geomorphic variability. The effects of inlets on coastal hydraulic and sedimentation patterns may extend to areas lying some distance from the inlet itself. Historical studies have shown that barrier inlets are ephemeral features, which may be closed or created at susceptible places by storm overwash during the course of a single storm. Newly breached inlets may be temporary or may be maintained by tidal currents and may persist for many years. Inlet shoals also are not fixed, and changes in form and dimensions occur as a result of varying currents, waves and sediment supply factors.

Barrier inlets and associated shoals are of great importance in barrier sedimentation because they act as sinks for sediment in the longshore transport system. The opening of an inlet can thus greatly reduce downdrift sediment supply by trapping large amounts of sediment moving in the longshore drift pattern and storing it in inlet shoal complexes. This loss of nourishment to downdrift beaches can result in serious erosion and shoreline regression as the sediment supply decreases. Inlet closure can restore downdrift sediment supply and ebb-tidal delta materials may gradually return to the alongshore

transport paths, but the flood-tidal delta materials remain virtually undisturbed in the lee of the restored barrier.

Many barrier inlets are important navigation channels, affording access to back-barrier lagoon complexes from the open sea. For this reason, it has become common practice to stabilize inlets and thus prevent migration and reduce shoaling of main channels with sediment. The chief means of stabilization has been the building of jetties and structural stabilization of the banks of the inlet throat. Since the jetties partly or wholly block alongshore transport, they may effectively trap some sediments that would formerly have migrated across the ebb-tidal shoals and reached the downdrift shore. Thus, jetties may create serious sand starvation in downdrift areas.

In recent years, various methods have been adopted to artificially bypass sand in inlet areas to prevent its loss into the inlet-associated shoals or its accretion on the updrift side of jetties. All require a detailed knowledge of local sediment transport and rates and pathways to materially affect inlet navigation, yet emulate the natural sediment transport processes.

Shelf Shoals

Large ridge-like shoals with a relief of up to 10 m or more that extend for tens of kilometers are common features of the continental shelf. They are especially well-developed and numerous in the Middle Atlantic Bight region, where extensive shoal fields occupy much of the shelf area. Similar shoals have been described on the shelf off Argentina and in the North Sea off Germany (Swift et al. 1978). Other shoals occur on the Mississippi River delta plain (Penland et al. 1989).

Shoals have attracted considerable attention because of questions regarding their origin and development and because most that have been investigated contain large amounts of clean fine- to coarse-grained sand and gravel potentially useful for beach fill or construction aggregate (Anders and Hansen 1990). Only a small number have been investigated in any detail as yet, using methods such as seismic reflection, profiling, and coring (Duane et al. 1972; Ludwick 1975; Coleman, Berquist, and Hobbs 1988; Meisburger and Duane 1971; Meisburger and Field 1975; Meisburger and Williams 1980; 1982). Much remains to be learned about the origin, nature, sediment characteristics, and variability of shelf shoals.

Linear shoals on the Atlantic Shelf exhibit at least 3 m of relief between the crest and the surrounding surface. Two types are recognized: shoreface-connected and isolated (Duane et al. 1972) (Figure 22). Shoreface-connected shoals show a seaward excursion of the 10-m depth contour, and isolated shoals occur farther seaward on the shelf floor. Isolated linear shoals are thought to have originated as shoreface-connected shoals during the Holocene

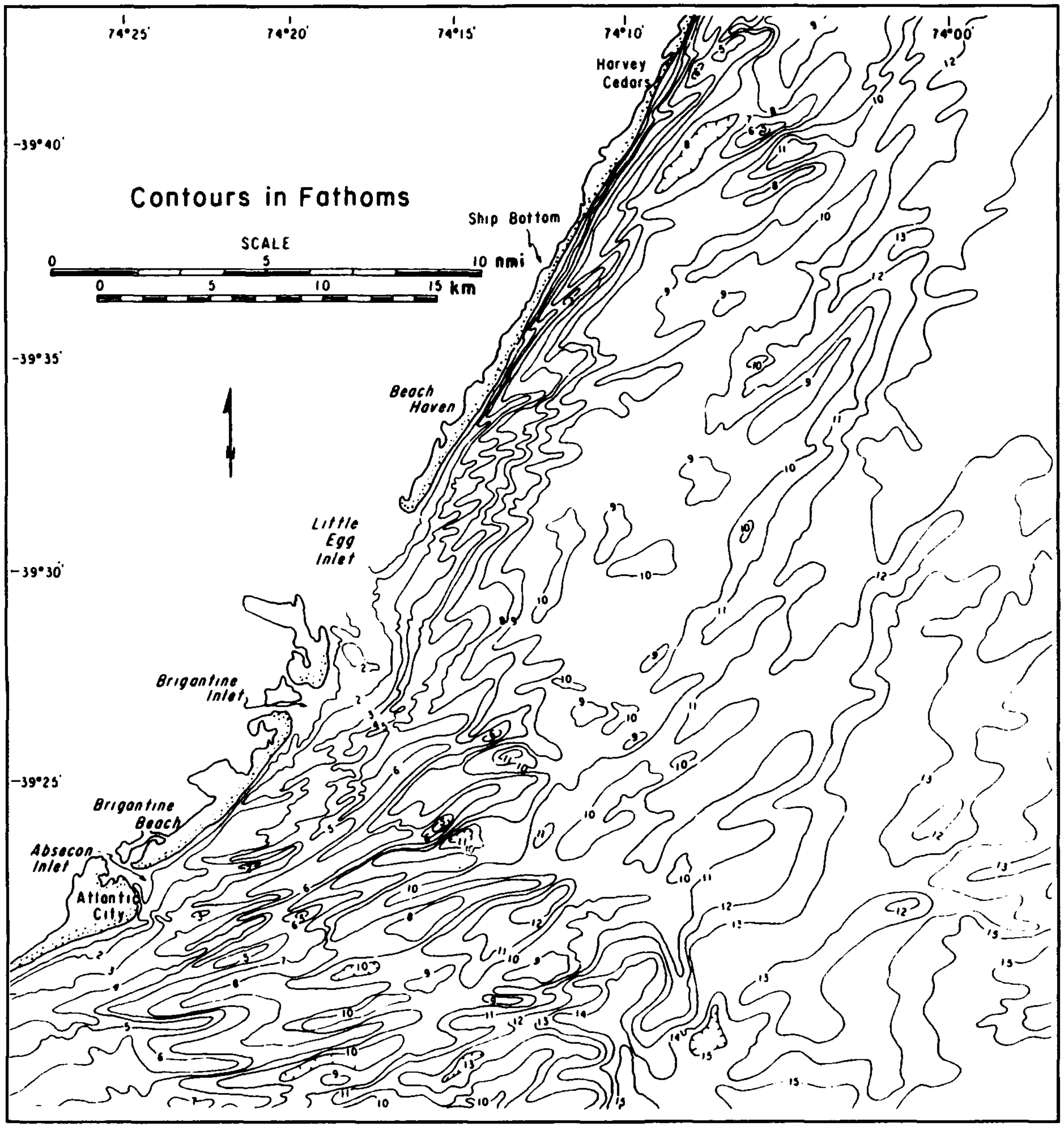


Figure 22. Linear shoals on the New Jersey coast. Note the abundance of shoreface-connected shoals throughout and the isolated shoals which are most abundant off the Brigantine area (from Meisburger and Williams 1982)

transgression that were detached by coastal retreat as sea level rose and flooded the shelf. Though nearshore processes were responsible for their formation, isolated shoals are maintained and modified by storm-generated shelf currents and waves.

Over 200 large shoals have been identified on the Atlantic Shelf, showing modal depths over the crest of 6 to 9 m, 12 to 15 m, and 24 m, mean shoal axis azimuths of 32 deg, and northward-opening angles of 10 to 35 deg between the shoreline and shoal axis (Duane et al. 1972). Seismic reflection profiles indicate that these shoals rest on a near-horizontal surface of the shelf floor. Cores of shoal material are primarily composed of mud-free, well-sorted medium sand or sand and gravel mixtures. In many cases, cores penetrating the shoal base have recovered mud, peat, and/or shells having a radiocarbon age of early to mid-Holocene.

Arcuate-shaped shoals, which occur in coastal areas, may be associated with estuaries or capes (Duane et al. 1972). Shoals may form near the mouths of estuaries and sounds where littoral drift material is intercepted and distributed by tidal currents. Cape-associated shoals (Figure 23) occur off cuspate forelands where convergence of littoral drift has resulted in the deposit of low-lying salients which protrude from the coast. Remnants of both estuary- and cape-associated shoals that were formed and abandoned during the Holocene transgression may extend well out on the shelf. These features in aggregate have been called shoal retreat massifs (Swift et al. 1972). They can be traced shoreward to modern arcuate shoals in estuaries and off capes.

The shapes and sizes of arcuate shoals and shoal retreat massifs are variable because of the influence of waves, and tidal and storm-generated currents. For example, Field and Duane (1974) found evidence that a shoal field off Canaveral Peninsula on the Florida Atlantic coast had been reworked by waves and currents. Sediments as deep as 4 m below the sediment-water interface showed evidence of recent abrasion. In addition, comparison of historical bathymetric surveys revealed that between 1898 and 1965, shoal crests accreted several meters and migrated over 300 m to the southeast. Similarly, studies of bathymetric charts of the past century by Granat and Ludwick (1980) indicate that shoals in Chesapeake Bay entrance changed shape and position during the time covered. Although relict in having originated in past times, most shoal retreat massifs have been influenced by subsequent shelf processes, often leading to radical changes in their form and orientation (Swift et al. 1972).

Deltas

Deltas are subaerial and subaqueous accumulations of river-derived sediments deposited at the coast. Sedimentation occurs when streams decelerate by entering and mixing with larger bodies of water (Wright 1982). Deltaic

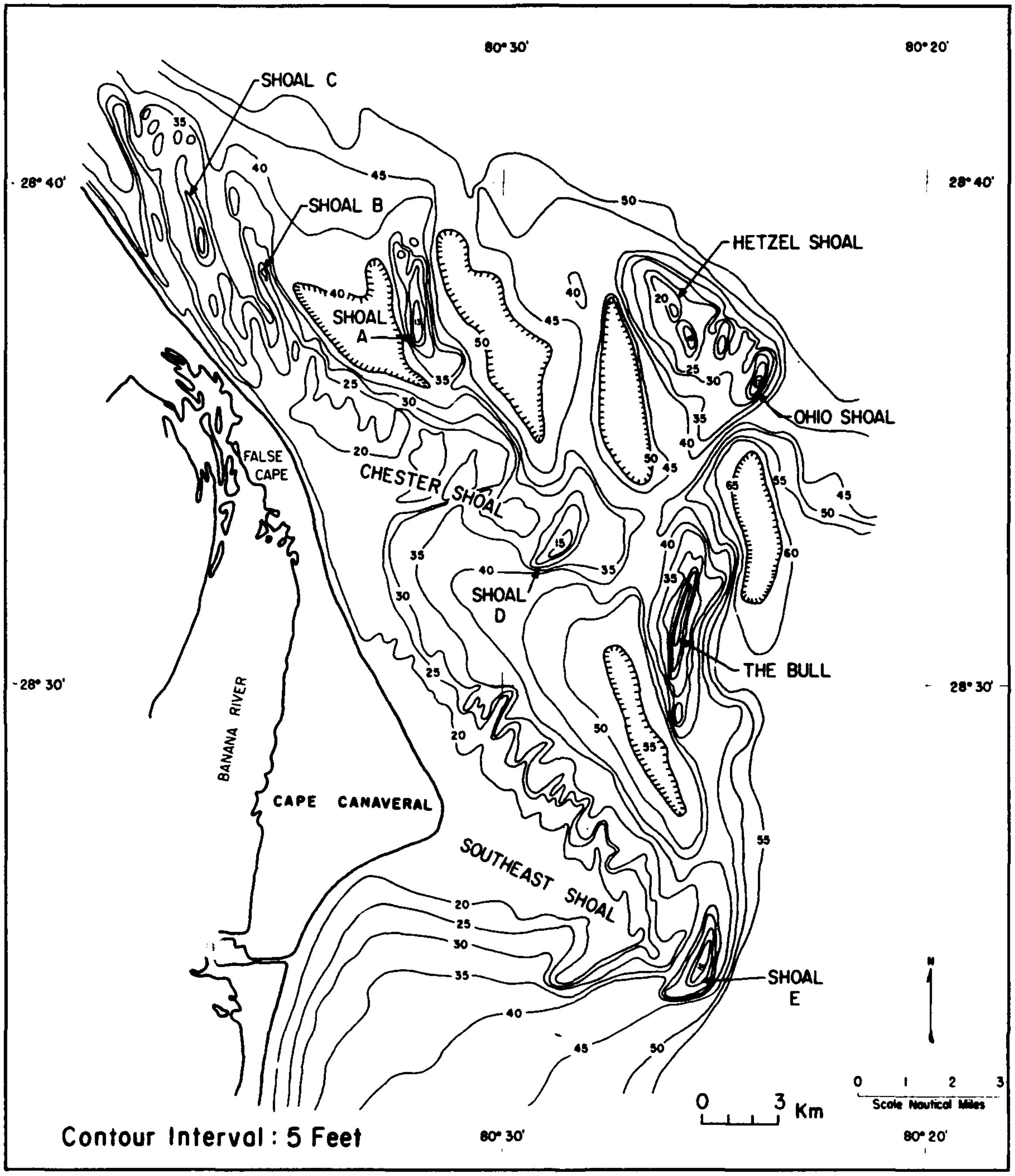


Figure 23. Cape-associated shoals off Canaveral Peninsula, a large cuspate foreland on the Florida Atlantic coast. Note shoreface-connected linear shoals north of False Cape (from Field and Duane 1974)

morphology, sediment distribution, and stratigraphy are the result of numerous factors related to the action and interaction of fluvial and marine or lacustrine processes and structural controls. Deltas develop when streams deliver more sediment to the coast than coastal processes can erode and transport away.

Deltas form under a wide variety of environmental conditions. A major prerequisite is the existence of a major river drainage system that carries substantial quantities of clastic sediment. Drainage basin climate, geology, relief, and area are all critical determinants of river discharge and sediment load. In addition to fluvial processes, marine processes (waves, tides, and currents) and tectonic and deformational forces influence delta formation and form. Though numerous factors tend to produce great variability in active deltas, it is generally agreed that the interaction of river, wave, and tide regimes is the major factor influencing delta morphology and sediment types (Galloway 1975). Littoral drift and offshore slope also control sand body geometry.

Subaqueous zones of deltas include the prodelta, the seawardmost element consisting of an apron of fine-grained sediments, and the more landward delta front, usually consisting of coarser silt and sand. The subaerial delta plain lies above the low tide level. It can be divided into a lower delta plain, where marine or lacustrine influence is active, and an upper delta plain, which lies above tidal or other marine or lacustrine influence and is dominated by riverine processes and features.

Active and abandoned zones occur on both the subaqueous and subaerial deltas. Active delta zones are accreting, and on the delta plain are occupied by functioning distributary channels, which develop from the deposition of bed material at channel mouths. Abandoned deltas occur where distributary channels became blocked or lost flow to other channels, causing them to fill with silts and clays. Abandonment can also result from upstream channel avulsion, in which the river occupies an alternative course with steeper gradient. Such events are attributed to seaward growth of the delta and the associated reduction in hydraulic head, a process which occurs about every 1,000 to 1,500 years in the Mississippi Delta Plain (Kolb and van Lopik 1966).

Depositional environments in deltaic areas can thus be distinguished according to which processes influence sedimentation. Although the nature of sediment supply may vary, the coarsest materials are deposited near the channel and mouth, while deposition of fine sediments occurs in environments of lower energy. The prodelta consists of fine-grained sediments deposited from suspension. On the delta front, features may include distributary mouth sands, distal-bar silts, tidal ridges, and shoreface beach deposits. On the delta plain, deposits associated with the distributary channel and margins include channel bed, channel fill, natural levee, and crevasse splay deposits. Interdistributary sediments are deposited in swamps, marshes, lakes, bays and lagoons, and are primarily laminated or bioturbated fine sediments. At the seaward edge, delta plain sediments of various kinds may be reworked into a

variety of environments more characteristic of coastal processes described elsewhere, including beaches, inlets, beach ridge plains or dune fields.

Once an active delta is abandoned, marine, estuarine, and paludal processes take over the landscape. Such abandoned deltas may prograde or remain stable if marine processes provide sufficient sediment. More likely, they may deteriorate because of marine reworking, increased subsidence and compaction inland, and decreased fluvial sedimentation rates. The Mississippi Delta Plain has undergone such transformations, where the sands of abandoned deltas have been reworked into barrier headlands and flanking barrier islands by marine processes. As the marsh behind these barriers subsides, barrier island arcs are left standing in open water. Through continued marine reworking and subsidence, the barrier islands progressively disintegrate to form inner shelf shoals. Transformation from one form to another may take several decades or hundreds of years, while delta abandonment to shoal formation may take a few thousand years to complete.

A delta plain, such as that of the Mississippi River, has unique environments that are highly vulnerable to both natural events and human activities. Because of storms such as tropical and extratropical cyclones, transgression of the low-profile barriers is rapid, with erosion rates over 5 m/year in places. In wetlands, many factors are involved in relative sea level rise and subsidence, including eustatic factors and consolidation of Quaternary and older sediments. In Louisiana, land loss has averaged over 100 sq km per year. Decreased sediment supply to wetlands because of levee building along the rivers and in the marshes, as well as decreased sediment loads due to trapping of sediments by reservoirs upstream, has prevented accretion of the marsh surface as sea level rises. Brine discharges, and dredging of canals and channels for navigation and oil-drilling activities are some other human activities that influence land loss.

Strand Plains: Ridges and Cheniers

On coasts where there is an abundant supply of unconsolidated sediments, and in wave-dominated settings, including in the vicinity of river mouths, it is common to find beach ridge plain coasts and chenier plain coasts. Both are called strand plains and display shore-parallel ridges, which represent successive seaward accretion and reworking (Figure 14). Ridges are shore-parallel bodies of coarse materials, with ridge crest elevations well above mean high tide and troughs near the mean low tide level. Ridges originate primarily due to marine processes, caused by several possible mechanisms. They generally develop immediately behind the active beach as a flood-level ridge, as an eolian accumulation, or by water deposition below and eolian deposition above.

Cheniers are shore-parallel bodies of sand and shell enclosed by prograding marsh and mudflat deposits. They develop in response to fluctuating supplies

of clastic sediment, typically in the vicinity of major river deltas. During episodes of reduced longshore sediment supply, coastal reworking occurs with deposition of coarse-grained material at the marine edge. During periods of accelerated fine-grained sediment supply, such as when delta lobes occur in proximity to these coasts, the coarse deposits are stranded inland behind seaward-building mudflat deposits.

4 Investigation of Environmental Factors

Geomorphic variability is chiefly caused by the work of dynamic environmental factors that vary over time and space. The most important of these factors are waves, tides, and currents which continually affect the shore and upper shoreface. During periodic storms, these factors affect a much wider zone, producing large-scale changes in geomorphic features. Because of this, data on geomorphic variations in the coastal zone are much more valuable when accompanied by wave and current observations for the same time period so that process-response relationships can be examined. This portion of the report concerns the equipment and techniques used to gather wave and current data.

Wave Data

Methods of obtaining wave data include gages, hindcasting from weather maps, shipboard observations, and littoral environment observations. Gages are the most accurate of these methods, but their relative cost often restricts their use to short-term deployments for the purpose of validating data collected by observation or hindcasting methods. Multiple gages across the shore zone in both shallow and deep water can be used to determine the accuracy of wave transformation calculations for a specific locale.

Wave gages can be separated into two general groups: directional and non-directional. In general, directional gages are more expensive to build, deploy, and maintain than non-directional gages. Because wave direction is an important parameter in applications such as sediment transport analysis and calculation of wave transformation, the additional expense is often necessary. Wave gages in both categories can be installed in buoys, placed directly on the sea or lake bottom, or mounted on existing structures, such as piers, navigation aids, and offshore platforms.

Buoy-mounted non-directional gages are accurate and relatively easy to deploy and maintain. Data are usually transmitted by radio from the buoy to

an onshore receiver and recorder. Bottom-mounted pressure gages measure wave parameters by sensing the pressure changes with the passage of each wave. They can be either self-recording or can be connected to onshore computers and recorders with cables. Divers must retrieve data periodically from self-recording gages. Both types of systems require regular maintenance. Structure-mounted wave gages are the most accessible of the non-directional gages, allowing convenient maintenance. Unfortunately, offshore structures are not always located near project sites.

Directional wave gages are used mainly in buoys or bottom mounts of single units (Figure 24) or multiple arrays in a fixed configuration. Directional buoy-type wave gages are often designed to measure other parameters, especially meteorologic ones. Pressure-type gages can measure wave direction using an accompanying electromagnetic current meter or by combining multiple synoptic pressure measurements from individual gages in a known geometric array.

Wave hindcasting is widely used for obtaining wave statistics by analysis of weather maps using techniques developed from theoretical considerations and empirical data (Coastal Engineering Research Center 1984) (Figure 25). Over the last several decades, since wave hindcasting came into common use, numerous improvements have been made in the technique and reasonably reliable information on wave climate in given areas can be computed (Abel et al. 1989; Hubertz and Brooks 1989; Jensen, Hubertz, and Payne 1989; Corson et al. 1987; Corson and Tracy 1987). Advantages of hindcasting include the long-term database associated with weather maps and the useful statistical information.

A large amount of wave data is available in the form of visual wave observations from ships at sea and from shore stations along the coasts of the United States. Although observations are less accurate than measured data, experienced persons can achieve reasonably accurate results and the large database of available observations makes it a valuable resource. Shipboard wave observations have been compiled by the US Navy Oceanographic Research and Development Activity in the form of sea and swell charts and data summaries such as the Summary of Shipboard Meteorological Observations. Areal coverage by these sources is extensive, but the greatest number of observations come from shipping lanes.

The second important source of observations has been collected by CERC under the Littoral Environmental Observation (LEO) program (Schneider 1981; Sherlock and Szuwalski 1987). The program, which was initiated in 1966, makes use of volunteer observers who make daily reports on conditions at specific sites along the coasts of the United States (Figure 26). A variety of data from over 200 observation sites are available from CERC. As shown, LEO data include more than wave parameters, and encompass information on winds, currents, and some morphologic features. LEO data are best applied to a specific site, and do not provide direct information regarding deepwater statistics.

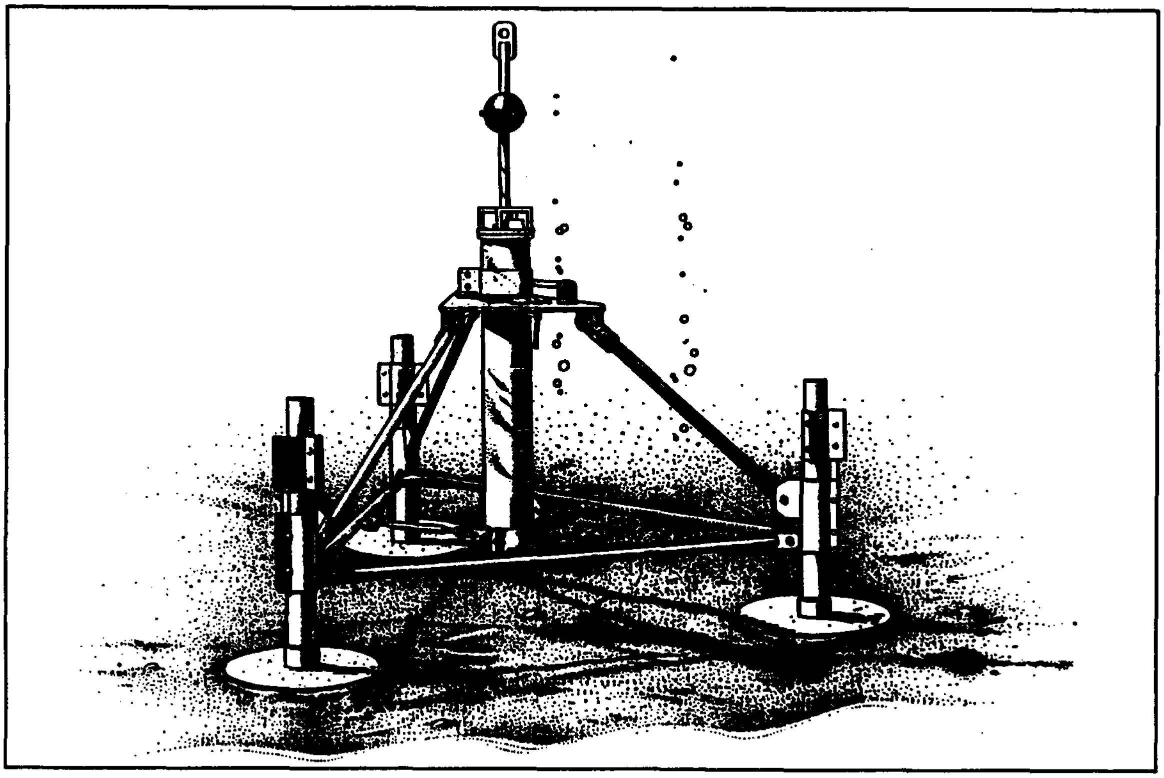


Figure 24. Bottom-mounted Sea Data™ 635-12 directional wave gage mounted in tripod using railroad wheels as corner weights

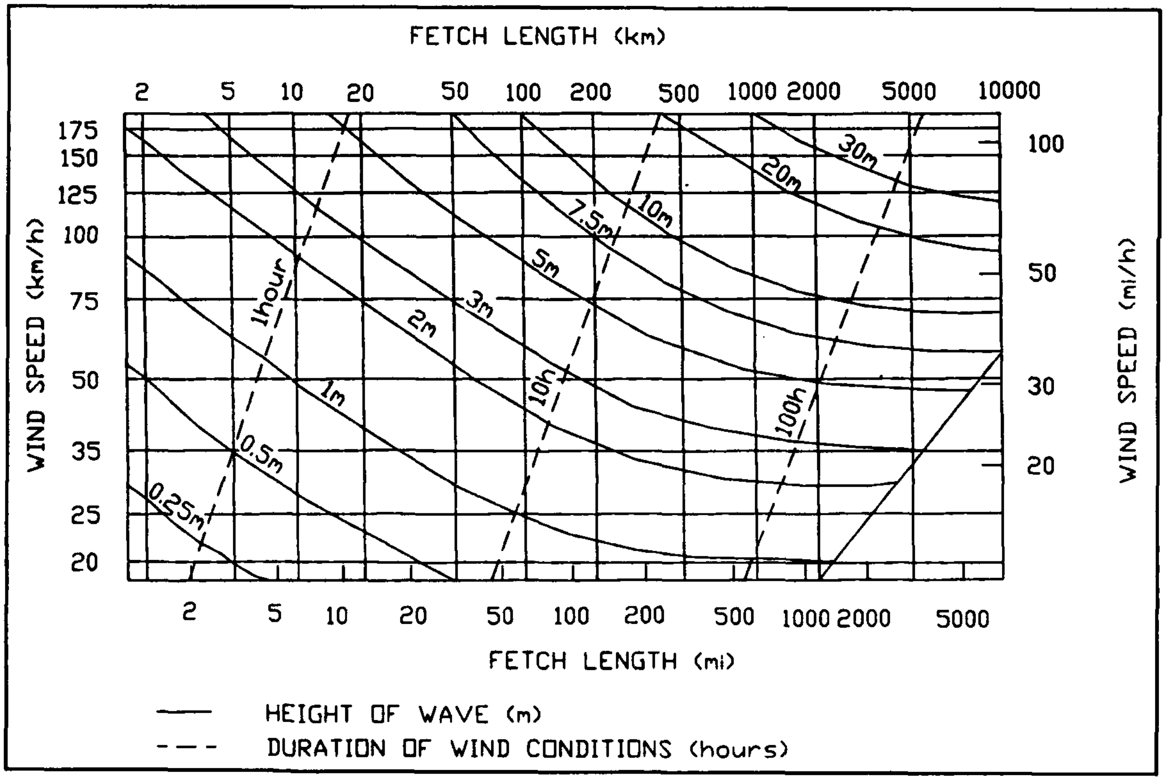


Figure 25. Deepwater wave hindcasting curves (from Bretschneider 1959)

LITTORAL ENVIRONMENT OBSERVATIONS

RECORD ALL DATA CAREFULLY AND LEGIBLY

SITE NUMBERS	YEAR	MONTH	DAY	TIME
1 2 3 4 5	6 7	8 9	10 11	Record time using the 24 hour system 12 13 14 15

WAVE PERIOD 16 17 18

Record the time in seconds for eleven (11) wave crests to pass a stationary point. If calm record 0.

BREAKER HEIGHT 19 20 . 21

Record the best estimate of the average wave height to the nearest tenth of a foot.

WAVE ANGLE AT BREAKER 22 23 24

Record to the nearest degree the direction the waves are coming from using the protractor on the reverse side. 0 if calm.

WAVE TYPE 25

0 – Calm
1 – Spilling
2 – Plunging
3 – Surging
4 – Spill / Plunge

WIND SPEED 26 27

Record wind speed to the nearest mph. If calm record 0.

WIND DIRECTION – Direction the wind is coming from. 28

1 – N 3 – E 5 – S 7 – W 0 – Calm
2 – NE 4 – SE 6 – SW 8 – NW

FORESHORE SLOPE 29 30

Record foreshore slope to the nearest degree.

WIDTH OF SURF ZONE 31 32 33 34

Estimate in feet the distance from shore to breakers, if calm record 0.

LONGSHORE CURRENT

DYE 36 37 38

Estimate distance in feet from shoreline to point of dye injection.

CURRENT SPEED 43 44 45

Measure in feet the distance the dye patch is observed to move during a one (1) minute period; If no longshore movement record 0.

CURRENT DIRECTION 46 47

0 No longshore movement
+1 Dye moves toward right
-1 Dye moves toward left

RIP CURRENTS 50 51 52

If rip currents are present, indicate spacing (feet). If spacing is irregular estimate average spacing. If no rips record 0.

BEACH CUSPS 54 55 56

If cusps are present, indicate spacing (feet). If spacing is irregular estimate average spacing. If no cusps record 0.

PLEASE PRINT:

SITE NAME ____________ OBSERVER ____________

Please Check The Form For Completeness

REMARKS: ____________

CERC 113-72
8 Mar 72

Make any additional remarks, computations or sketches on the reverse side of this form.

Figure 26. Littoral Environmental Observation forms used by the volunteer observers participating in the LEO program (from Schneider 1981)

While a considerable amount of data exist on the coastal and oceanic processes of the US coasts, there are few field studies in which data on geomorphic changes and relevant processes were obtained concurrently. Such data are critically needed for many areas in order to validate existing models of process/response relationships under a variety of conditions.

Wave data is one of several components required to characterize the process-response framework of the coastal zone. Important wave parameters include wave height, period, and steepness, and breaker type. The estimated height is often given as the significant wave height ($H_{1/3}$), the average of the highest one-third of waves, or as the maximum height ($H_{1/10}$), the average of the highest one-tenth of the waves. Significant wave height may be used to compute other wave statistics (Shore Protection Manual 1984).

Certain wave characteristics are strongly related to morphologic variables. Wave steepness, for example, is an important variable in determining foreshore slope, which explains changes in beach profile characteristics from summer to winter (Shepard and LaFond 1940; Saville 1950; Bascom 1954). Other wave characteristics, including the surf scaling factor and breaker type are important in determining beach profile characteristics (Wright et al. 1979; Huntley and Bowen 1975). Other parameters, such as shore-normal currents and sediment grain size, should also be considered in conjunction with wave variables in order to thoroughly understand beach profile development (Sonu and van Beek 1971; Iwagaki and Noda 1963; Komar 1976).

Wave climate data can also be used in conjunction with bathymetric data to construct wave refraction diagrams, which provide an indication of how bottom topography can affect the bending of waves approaching a shoreline. Such studies can help in determining mass transport and longshore transport of sediment, which in turn can assist in predicting morphologic changes, and in design of coastal engineering projects. Wave refraction analysis can also be used for hypothetical scenarios, for instance, how wave energy and associated littoral conditions would be affected by the dredging of an offshore shoal or offshore placement of dredged material.

Water Level

Water level measurements represent the combined effects of tides, setup or setdown by onshore or offshore winds, eustatic changes, and vertical crustal displacements Figures 27-31). Short-term variations, particularly those associated with storms, are important in increasing the effective wave base. This allows erosion to take place farther inland than during mild weather periods. Long-term variations in sea level, while of much lower intensity than surf processes, can be important in predicting erosion or accretion and changes in beach profile response (e.g., Wells and Coleman 1981b; Hands 1983).

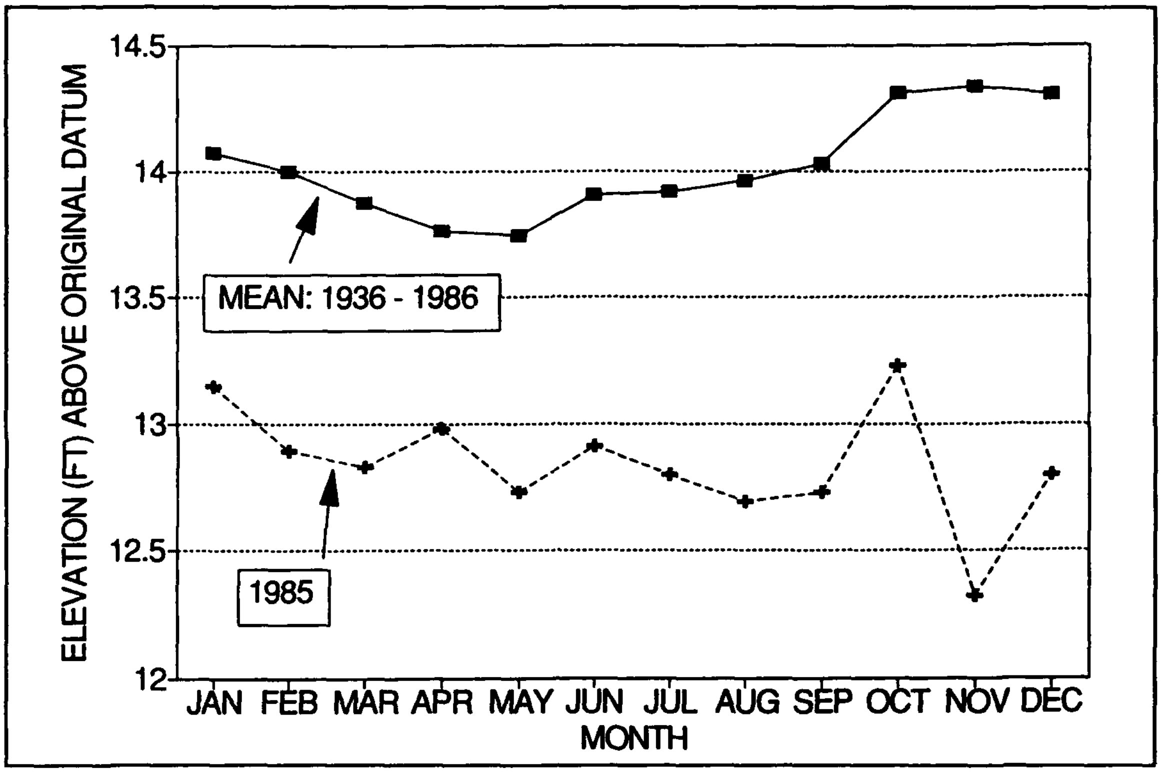


Figure 27. Monthly water level changes at Juneau, AK. High water typically occurs October-December. Data from Lyles, Hickman, and Debaugh (1988)

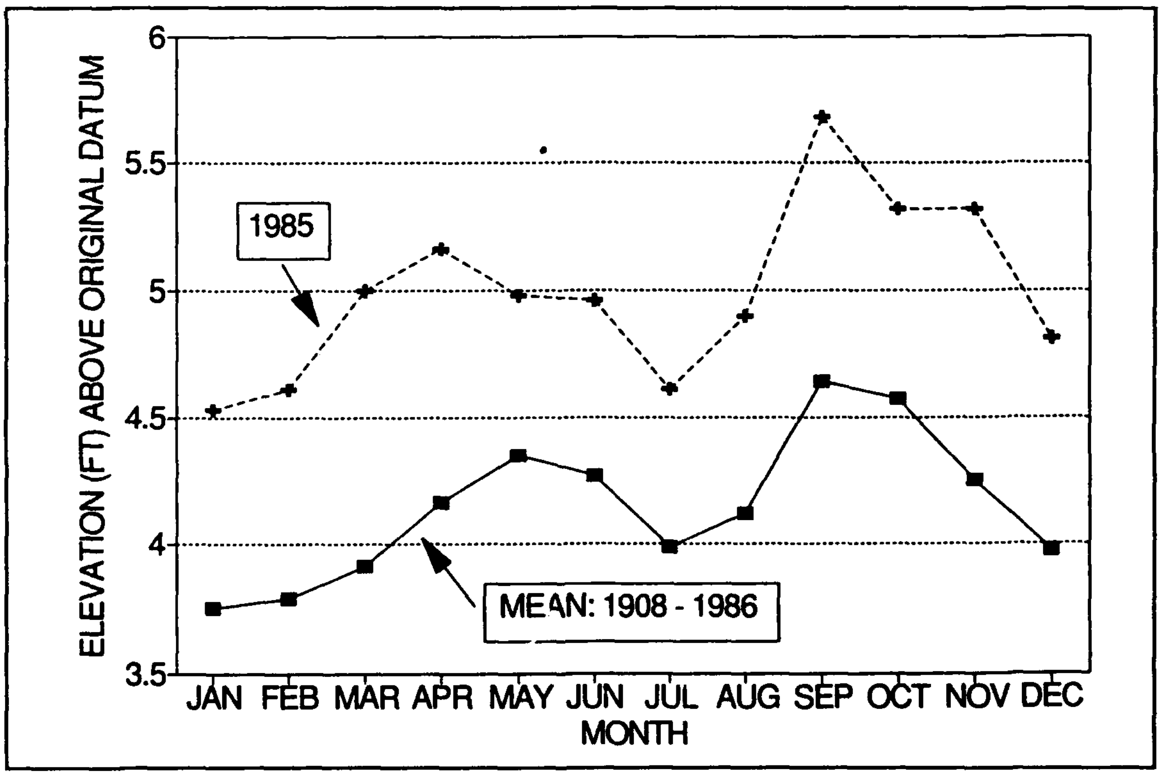


Figure 28. Monthly water level changes at Galveston, TX. High water occurs twice per year: April and September-November

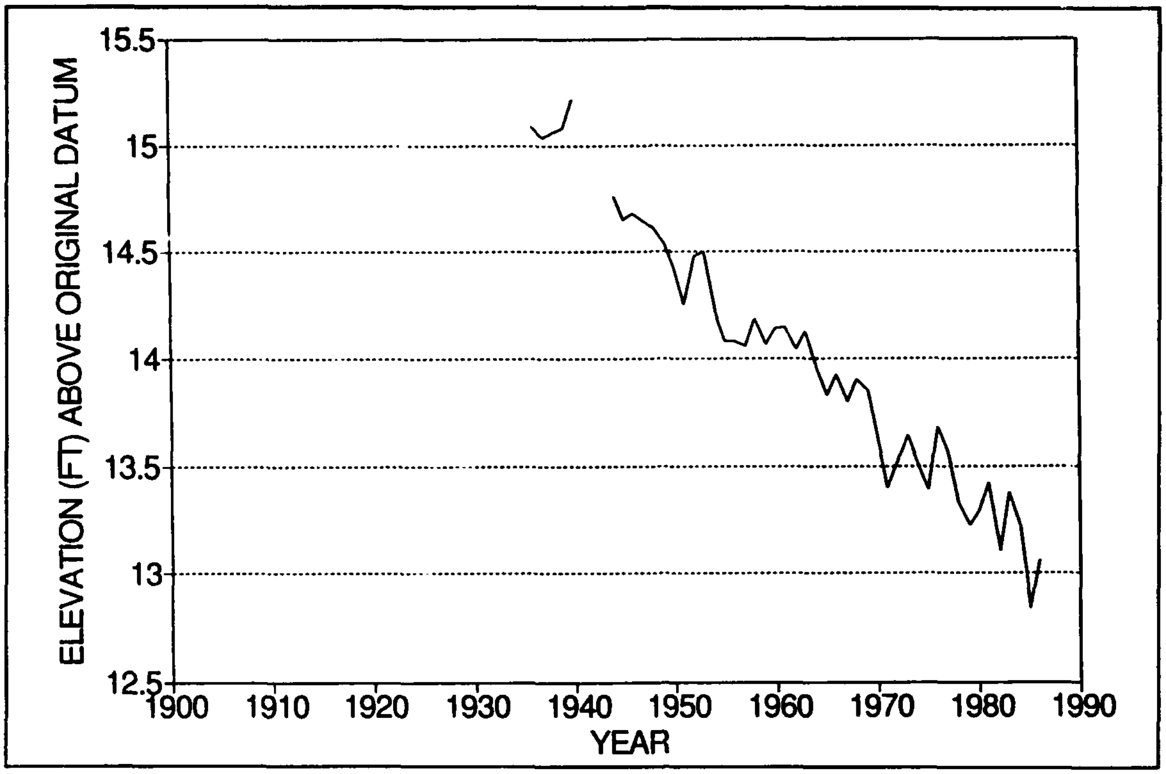


Figure 29. Yearly mean sea level changes at Juneau, AK, from 1936-1986. The overall fall in sea level shows the effects of isostatic rebound. Data from Lyles, Hickman, and Debaugh (1988)

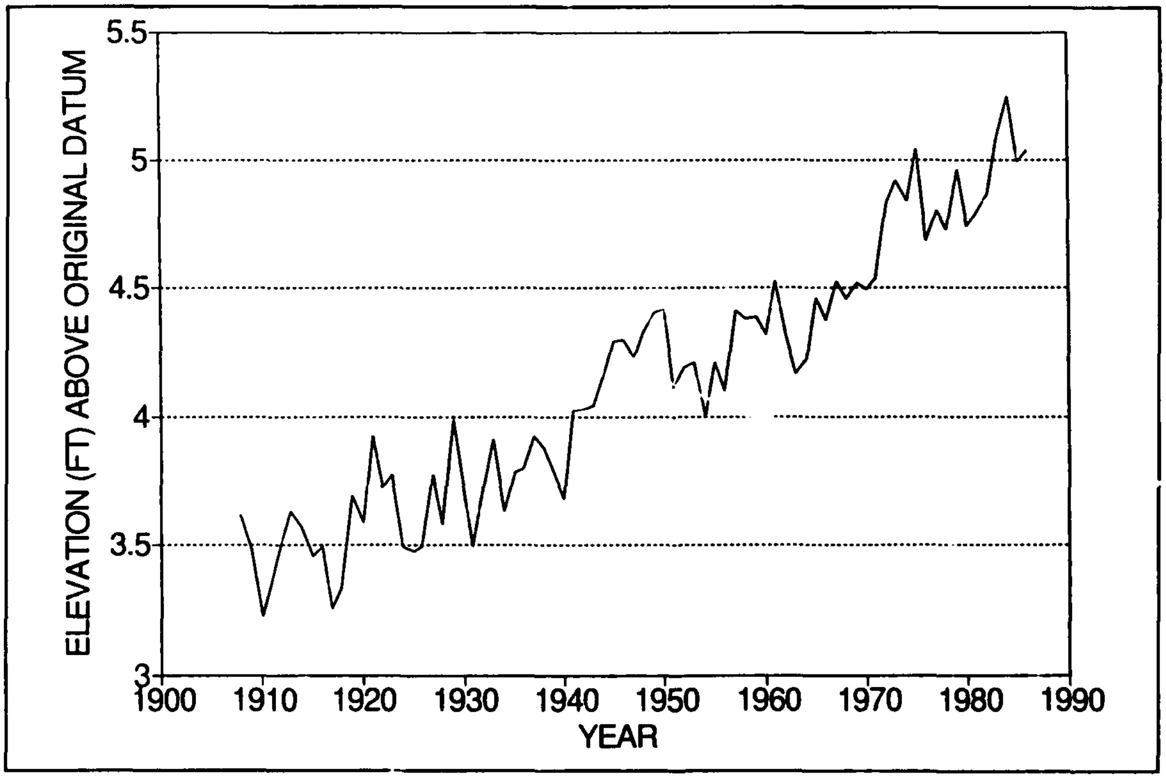


Figure 30. Yearly mean sea level changes at Galveston, T , from 1908-1986. Subsidence of the land around Galveston may be caused by groundwater withdrawal and compaction

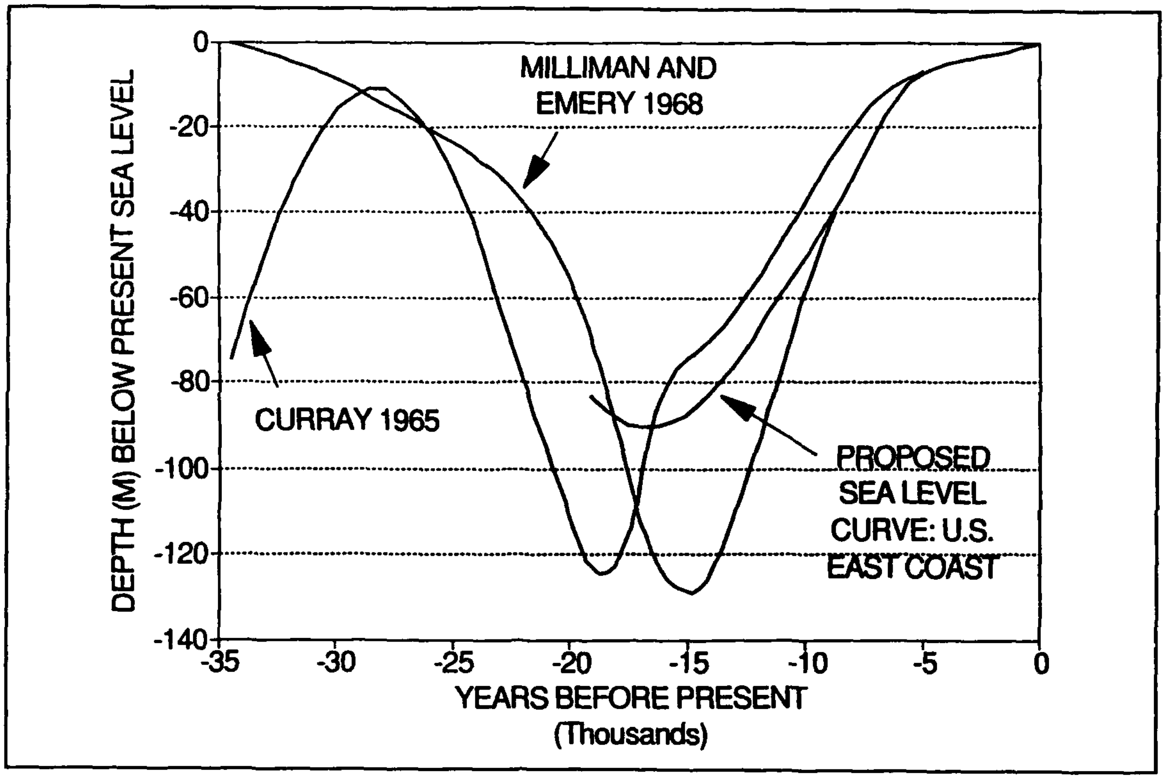


Figure 31. Late Quarternary sea level curves inferred from radiocarbon-dated samples along the U.S. coastlines (modified from Dillon and Oldale 1978)

Over shorter time scales, tides and storm surges are important in understanding geomorphic changes. In order to obtain continuous data at a specific site, water level recorders have to be deployed. If fewer data are required, high and low tidal predictions can be obtained from tide recording stations. However, in some locations, the discrepancy between predicted and actual tides on coastal areas only a short distance apart may be considerable. A method for tidal adjustment between predicted tides at a station and those at a nearby study area can be obtained even if only short-term site measurements are available (Glen 1979).

Sea level also shows pronounced seasonal changes, with some locations differing by 1 m annually from the highest to the lowest monthly values (Komar 1976). Around the United States and in most locations worldwide, for example, sea level is lowest in the spring months and highest during the late summer and autumn. A number of factors are responsible for these seasonal deviations, including changes in water temperature, salinity, atmospheric pressure, river runoff, and longshore winds.

The National Ocean Service of the National Oceanic and Atmospheric Administration (NOAA) is responsible for monitoring sea level variations at 115 stations nationwide (Hicks 1972b). The Corps of Engineers District offices located near the coast also collect tidal elevation data at additional locations. Daily readings are published in reports that are entitled "Stages and Discharges of the (insert location) District."

Trends of sea level series over decades have been examined for the United States and elsewhere, in some cases dating back to the beginning of the century (Hicks 1972b; 1978; Gornitz, Lebedeff, and Hansen 1982). In the analysis of long-term data, Hicks (1972b) found that series exhibited yearly variability and apparent secular trends, which might either be nonperiodic phenomena or segments of very long oscillations. Yearly variability is due to variations in the meteorological and oceanographic parameters of wind, direct atmospheric pressure, river discharge, currents, salinity, and water temperature. In one case, extreme unidirectional change was caused within a few months by the Alaskan earthquake in 1964 (Hicks 1972a). Apparent secular trends result from longer term glacioeustatic, tectonic, climatologic, and oceanographic influences.

In some cases, it is important to dampen the effects of yearly variability so that the nature of secular trends will become more pronounced. A weighting array may sometimes be applied to reduce yearly effects. Least-squares regression methods are typically inadequate, as the secular trends often show pronounced nonlinearity (Hicks 1972b). It is also important to examine periodic effects in the long-term series, such as the 18.6-year nodal period, which Wells and Coleman (1981b) concluded was important for mudflat stabilization in Surinam. In some cases, such as on the Great Lakes, water level changes in conjunction with existing models can be used to predict erosion and changes in the shore and offshore profile (Hands 1983).

The geomorphic response of the shoreline to sea level rise may follow one of several scenarios. Applicability of the erosional response model or Bruun Rule has been discussed by Hands (1983). The onshore migration or rollover model often applies to barrier coasts where washover is an important process (Dillon 1970). In other locations where coastal forms change slowly, features may drown in place without transport occurring (Carter 1988). Depending on the location and time scale involved, the geomorphic effects can be studied with a variety of techniques including historic data, seismic data, and stratigraphic methods.

Currents

Speed and direction of longshore and cross-shore currents are also important for understanding coastal changes. Direct measurements of the velocity and direction of current flow can be made by instruments deployed on the bottom or at various heights in the water column. Lagrangian methods such as floats, bottom drifters, drogues, and dye are also used, especially in the littoral zone where current meter data (as well as the current meters) are adversely affected by turbulence.

Current measurements may be Lagrangian, following the motion of the flow in its spatial and temporal evolution, or Eulerian, defining the motion at

a fixed point and determining its temporal evolution. Lagrangian current-measuring devices are often used in circulation studies, pollution studies, and for monitoring ice drift. For Eulerian or fixed current meters, proper placement is essential for adequately determining sediment transport pathways.

Several types of current sensors are in common use including impeller, electromagnetic, acoustic, acoustic Doppler, laser Doppler, and inclinometer types (Fredette et al. 1990). Impeller current measurements are acquired by means of a propeller device which is rotated by current flow. Impeller devices are considered to be the least expensive and have been widely used for a considerable time (Teleki, Musialowski, and Prins 1976). They are subject to snaring, biofouling, and bearing failure, but are more easily repaired in the field and more easily calibrated than other types (Fredette et al. 1990).

All the other current meters have several features in common, although they operate on different principles. Each has no moving parts, has rapid response, is self-contained, can be used in real-time systems, and can be used to measure at least two velocity components. The degree of experience of the persons working with the instruments probably has more to do with the performance of the current meters than does the type of meter used (Fredette et al. 1990).

In addition to direct current measurements, indirect current estimates of current speed and direction can also be made from bedforms, particularly in shallow water. The deviations from a flat bed and associated sedimentary structures are associated with coastal hydrodynamics, including the effects and interaction among tidal currents, wave characteristics, and longshore drift, particularly at inlets and estuaries. Bedforms reflect flow velocity, but are generally independent of depth (Clifton and Dingler 1984; Boothroyd 1985). Bedform can vary in response to increasing flow strength (Hayes and Kana 1976). Bedform orientation and associated slipfaces also provide clues to flow direction.

Knowledge of the magnitude and direction of currents at the coast allows the prediction of sediment movements and thus is basic to an understanding of landform development. Information concerning cross-shore (shore-normal) currents and sediment transport can assist in predicting beach profile change. Longshore (shore-parallel) currents and associated transport can assist in predicting beach planview changes. The combined effects of both types of currents, generating cell circulation, may explain or assist in identifying regularly spaced features along many coasts. The longshore migration of such cells can also cause landform migration associated with the spatiotemporal migration of higher energy nodes.

Conversely, the configurations of the shoreline can provide information regarding littoral currents. Shoreline proturberances, particularly in the vicinity of structures, headlands and barriers, and tidal inlets are useful indicators of the prevailing littoral sediment drift (Komar 1976). Such indicators cannot generally be used for quantitative estimates of the sediment

transport rate because the shoreline eventully tends to reach an equilibrium condition.

Some types of currents, such as rip currents, may be controlled in spacing by other parameters; i.e., edge waves or other wave height variations along the shore, or the surf zone width. However, irregular nearshore topography, which also may be manifest by shoreline protuberances, can produce nearshore circulation (Sonu 1972). In such cases wave height variations may not exist. Aerial photography may be helpful in assessing the location of some types of currents, their patterns, and possibly their movement. Locations of rip currents can sometimes be detected using side-scan sonar if characteristic channels have been scoured in the seafloor (Morang and McMaster 1980).

5 Investigation of Geomorphic Factors

Nature of Geomorphic Changes

Techniques for investigating geomorphic features and coastal evolution can provide useful information for coastal engineering design. Such techniques can include field surveys, analysis of historical maps and aerial photographs, airborne remote sensing, waterborne remote sensing, and sedimentologic and stratigraphic techniques depending on the spatial scale and the time scale of the data needed. Collection and comparative analysis of time series data showing dimensions, elevations, and configuration of coastal features over an extended period of time are effective ways of identifying temporal geomorphic changes and trends. In most cases, the use of multiple techniques will provide a useful balance of information regarding a site.

At the interface where marine and lacustrine processes interact with the land, coastal features are highly variable. For effective management, cyclic patterns, intermittent noncyclic events, and long-term trends must be considered. In mild weather, the morphological changes that take place on a day-to-day basis are relatively small and often compensate for each other, so that little net effect is apparent. During storms, a wider zone of the coast may be exposed to coastal processes and large geomorphic changes can occur in only hours or days.

In many places, distinct changes in coastal geomorphic features occur on a seasonal basis. A common example is a winter-summer cycle, in which winters are times of more intense and frequent storms than the summer seasons. An example of seasonal cycles in coastal morphology is the changes that occur in beach profiles. The more severe wave climate of winter causes erosion of the shore, with the eroded material usually transported seaward to the upper shoreface, where it often forms submarine bars. With the return of milder conditions in the summer months, this sand usually returns to the beach and a period of milder weather prevails. This cycle may be interrupted by hurricanes in summer and early fall, which greatly disrupt the normal summer

shore characteristics and often penetrate well inland of the shore zone because of high waves and storm surges.

In addition to cyclic seasonal patterns, many areas are being affected by long-term unidirectional trends in a particular environmental factor, which causes continuous adjustments in morphology. For example, a rising relative sea level usually results in shore erosion and progressive landward retreat of the shoreline. Evidence of past or ongoing changes, and the rates of these changes, need to be taken into account for planning and management of engineering activities.

A rapid change in coastal morphology, due to an intense storm or other event that causes a feature to be out of equilibrium with the prevailing environmental factors can lead to significant long- and short-term trends. Unidirectional trends are observed until the normal balance or equilibrium is restored. In some cases, the effects of an extreme event may become permanent because normal processes are unable to restore the old equilibrium.

Several types of unidirectional trends caused by extreme events are possible. For example, beach and dune sands lost to storm overwash deposits cannot be returned because there is no process that can move the material back toward the shore, except for minor eolian transport. Thus, most of the material will not reenter the shore area unless barrier recession eventually reexposes it on the seaward side. Also, if island breaching occurs during storms, inlets may develop and enlarge unless littoral drift processes work to close them.

The identification and analysis of trends in coastal geomorphic processes and features are of great importance in the planning and design of coastal engineering efforts and of long-range management plans. Indeed, in many cases, the purpose of coastal engineering works may be to modify or compensate for some trend that is producing undesirable effects. Since each location is unique geomorphically, exhibiting different types of changes, management and structural techniques that are appropriate for a given site might not be appropriate for others. In summary, although coastal geomorphology displays many general trends, unique conditions at each location must be identified and evaluated before initiating engineering projects or long-term management practices.

Historical Charts and Aerial Photographs

Detection of long-term trends is often difficult because these processes are often relatively slow on human time scales. However, trends may be identified by comparative analysis of historical maps, charts, and aerial photographs that show changes over a period of decades or centuries. Historical charts and aerial photographs of many areas have been periodically

resurveyed and fairly accurate surveys, going back in some cases 150 years or more, are available for most US coastal areas.

The primary source of chart data is the NOS and its predecessor, the U.S. Coast and Geodetic Survey. Archive material for all past surveys of these agencies is available from NOS, a division within NOAA. Much of this data can be obtained in the form of preliminary plots that are of a larger scale and contain more soundings and bottom notations than the published charts made from them. Some of these data were used for regional studies of net shoreline movement (Anders, Reed, and Meisburger 1990) (Figure 32). A great variety of additional documentary evidence may also be available, as described by Fulton (1981). This includes such items as local records and tax assessments, which might be incorporated in background investigations of a site if sufficient time exists to find these materials.

Aerial photographs are another useful and economic technique for examining details of coastal features above the water line. The general turbidity of coastal waters inhibits the application of photographic data of the offshore bottom; however, in relatively clear shallow water, the crests of submarine bars and shoals may be visible. Sources of aerial photography data are numerous, including Federal, state, county, and local government agencies. Aerial photography of coastal areas has been collected for about 60 years; thus, it often can provide useful time series data on changing conditions. Also, the effects of major events can be documented by aerial photography because the necessary equipment and airplanes can be rapidly mobilized to reach areas that are not easily accessible on the ground.

Applications of aerial photographs include assessments of short-term and long-term, as well as mesoscale and macroscale, coastal changes. The type of information that can be derived depends in part upon the scale of the photography, and also upon the historical nature of the database. The relative accuracy of the surveys, maps, or aerial photographs will depend largely on the scale of the initial photography; horizontal and vertical error increases with smaller scale (Tanner 1978).

Much information regarding local processes can be derived from maps and aerial photographs. One example is the longshore movement of sediment, an item of paramount interest to geologists and engineers because of its importance in coastline evolution. The geometry of coastlines in the vicinity of headlands, tidal inlets and streams, and coastal structures is one key to determine the directions of littoral transport (Figure 33). Storm impacts, including island breaches, occurrence of overwash features, and changes in inlets, vegetation, and dunes can be determined with time-series photography. Problems with siltation of tidal inlets, river mouths, estuaries, and harbors can also be examined using photographs.

Large data sets of historical aerial photographs and maps can be used for interpretating regional geomorphic changes in coasts. Using detailed historical data, Dolan and Hayden (1983) were able to assess that shore processes and

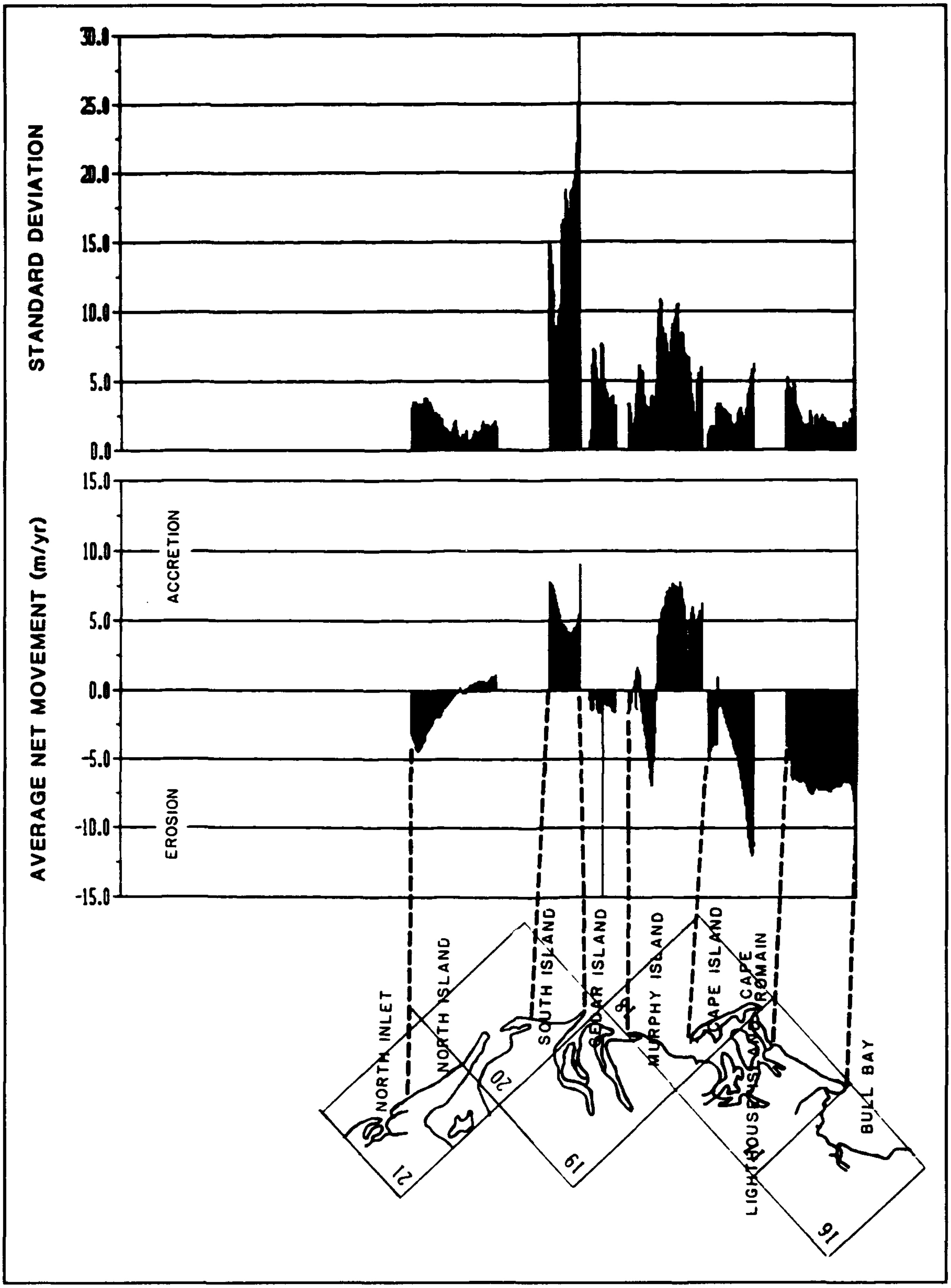


Figure 32. Net shoreline movement of a portion of the South Carolina coast based on historical maps surveyed between 1857 and 1983 (from Anders, Reed, and Meisburger 1990)

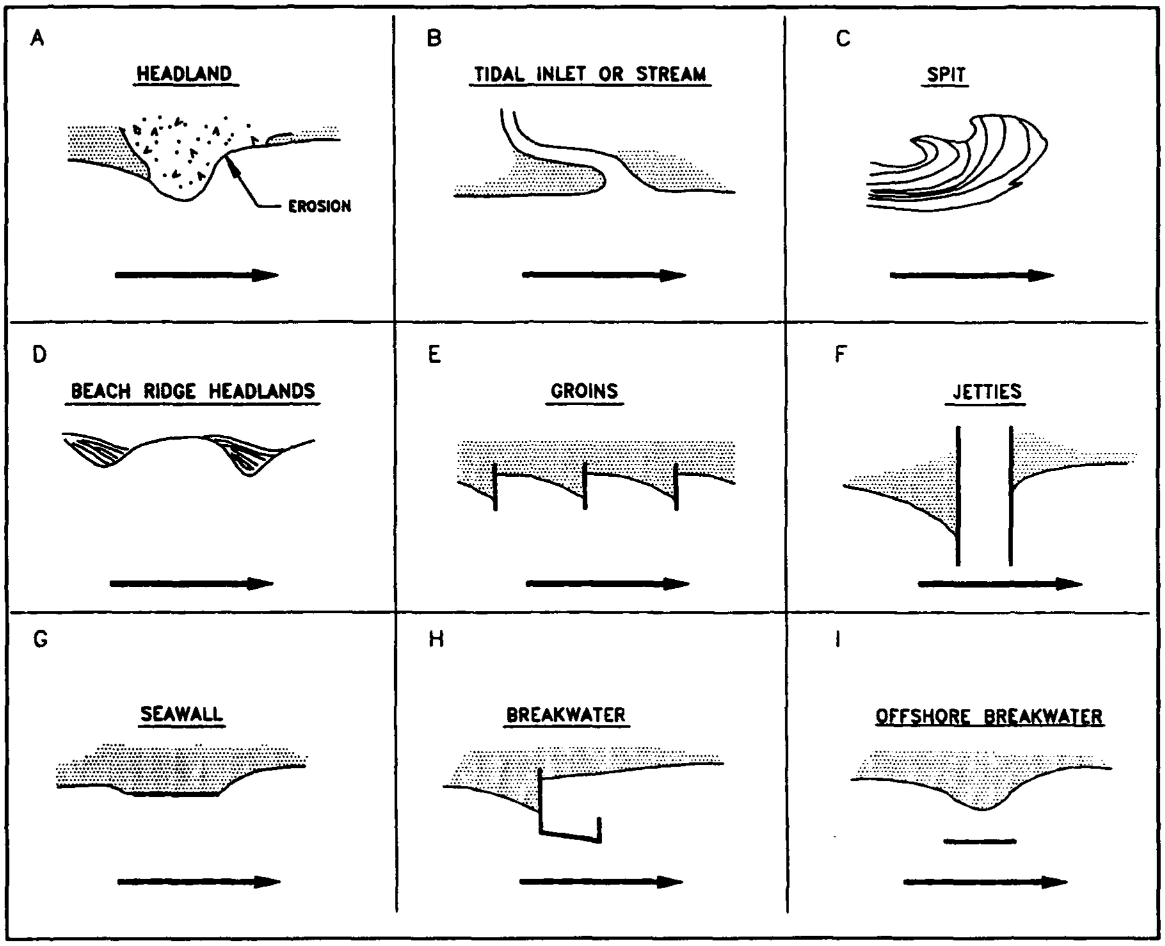


Figure 33. Morphologic indicators of littoral drift along natural and modified shorelines. Natural features such as rock headlands show accretion on the updrift side and erosion on the downdrift side (A), tidal inlets and spits show extension in a downdrift direction (B-C), and beach ridge headlands show successive growth on the updrift end influenced by the development of coastal cells, which form shoreline irregularities (D). Coastal engineering structures including groin fields, jetties, seawalls, attached breakwaters, and detached breakwaters (E-I) generally show accumulation of sediment on the updrift side, and reduced sediment supply on the downdrift side

landforms assume systematic, as opposed to random, patterns both along and across the coast. Large storms caused severe erosion in the same locations as previous storms of lower severity. Long-term erosion rates can be examined even over large areas (Dolan, Hayden, and May 1983). Such studies can then be used to assign temporal probability levels to the distribution of coastal change rates and thus predict shoreline positions several years beyond the data set (Dolan et al. 1982; Dolan and Hayden 1983). Clues regarding the influence of natural and human-induced wetland changes can also be found by interpretation of historical maps and photographs (May and Britsch 1987).

If detailed historical mapping of shorelines or inland areas is attempted, care should be taken to reduce photographic distortion error. Aerial photography should be corrected photogrammetrically to reduce sources of error such as tilt, yaw, and parallax. If data are taken from maps using different projections, the maps need to be geometrically corrected to conform to the same grid. Once these corrections are made, allowance should be made for tidal and seasonal changes. Good reference points should be chosen. These are usually readily available in populated areas or areas with strong human imprint, but can be difficult to establish along non-populated coasts. Sampling intervals must also be chosen to avoid systematic errors associated with rhythmic features. Extrapolation errors should also be avoided, so that long-term erosion rates are not projected from short-term data sets.

Airborne Scanners and Satellite Data

Multispectral scanners and remote sensing devices mounted on aircraft and satellites can provide various types of data that in some cases exceed the capability of conventional photography. The resolution of scanner data is generally not as good as aerial photography, with each data cell or pixel (picture element) having a size from a few meters (on low-altitude aircraft scanners) to several hundred meters on a side. Depending on the system and flight altitude, the aerial coverage on one image is typically far more extensive than on photographs. Also, the coverage may extend to parts of the electromagnetic spectrum invisible to the human eye, including the near and thermal infrared and radar bands. With digital data collected from scanners, the capability for quantitative analysis is far superior to aerial photography, and the data in some cases may be suited to numerical modelling studies. Time-sampling capabilities at a given location include several passes per month by different satellites, allowing repetitive changes to be examined. The timing, however, is not at the discretion of the user.

Applications of satellite remote sensing are especially good for assessing large-scale changes in the surface of the coastal zone. In the vicinity of deltas, estuaries, and other sediment-laden locations, the determination of spatial patterns of suspended sediment concentration can be facilitated with remote sensing. In shallow-water portions of non-turbid water bodies, some features of the bottom, including the crests of submarine bars and shoals, can be imaged. On a relatively crude level, satellites may assist in monitoring tidal changes, particularly where the land-sea boundary changes by several hundreds of meters. In deeper waters, satellites can also provide data on ocean currents and circulation (Barrick, Evans, and Weber 1977). Aircraft-mounted radar data also show promise in the analysis of sea state.

Remote Sensing By Ocean Vessel

Depth-sounders, side-scan sonar, and subbottom profilers are three major ways to collect data on subaqueous environments from an ocean vessel. All require the use of high resolution positioning systems. Fathometers are the most common devices used for acoustic depth-sounding to conduct bathymetric surveys. Side-scan sonar provides an image of the aereal distribution of sediment, surface bedforms, and larger features such as shoals and channels, and thus can be helpful in mapping directions of sediment motion. Subbottom profilers and seismic techniques are used to examine the near-surface stratigraphy below the seafloor.

Bathymetric surveys are required for many studies of geomorphic variability in coastal waters. Survey-quality fathometers include many devices to improve accuracy, including correction for change in tide, change in draft with vessel speed, and speed of sound. Even so, the maximum accuracy is estimated to be $\pm$ 0.2 m (Morton, Stewart, and Germano 1984); thus, such errors should be considered in volume change measurements of features. Survey lines are typically parallel and are run at an appropriate spacing depending on the survey's purpose and the scale of features to be examined.

Side-scan sonar permits the collection of surface characteristics data for the seafloor. The resulting image of the bottom is similar to a continuous, oblique aerial photograph, with lower frequency scanners providing less detail but greater range than higher frequency scanners. Detailed information, including the spacing and orientation of bedforms, and grain size differences in seafloor sediments, can generally be distinguished on side-scan, as can larger individual features. It is generally recommended that bathymetry be run in conjunction with side-scan, because detailed information on the relief of bottom features is valuable during the interpretation of side-scan sonographs. The side-scan system is sensitive to vessel motion, making it suitable only for work during calm conditions.

The principles of subbottom seismic profiling are fundamentally the same as in acoustic depth sounding. Subbottom seismics employ a lower frequency, higher power signal to penetrate the seafloor. The signal is reflected from interfaces between sediment layers of different acoustical impedance (Figure 34). Coarse sand and gravel are often difficult to penetrate with conventional subbottom profilers, resulting in poor records. New equipment is helping to overcome such problems, although the data are of lower resolution. Spacing and grid dimensions are usually the same as those used for bathymetric and side-scan surveys, being dependent upon the nature of the investigation.

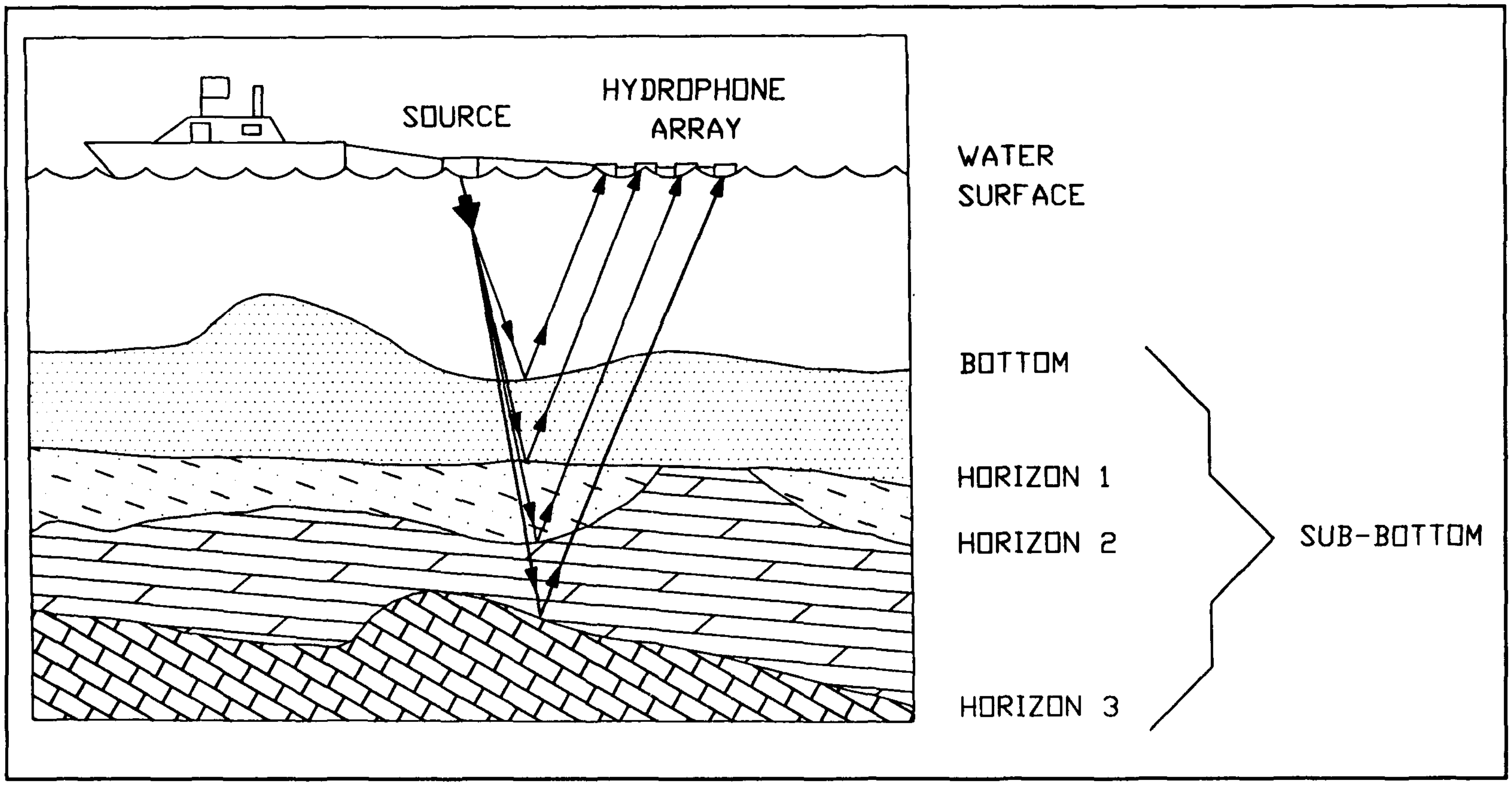


Figure 34. Principles of obtaining subbottom seismic data

Field Survey Techniques

The most direct and accurate means of assessing geomorphic variability is to conduct periodic surveys for the express purpose of obtaining time-series data. However, as a practical matter it is usually not feasible to carry out repeated surveys for a sufficient length of time for reliable and comprehensive information because of expense and because the lead time for projects often does not allow sufficient time to obtain the needed data. Nonetheless, a set of surveys spanning over a year or more can be of substantial help in learning more about the prevailing seasonal changes.

Field surveys of coastal features show that the most active zones are the shore and upper shoreface. Submerged interior parts of previously subaerial features, such as relict shore and dune deposits formed during earlier stages of development, are distant from the modern shoreline and are likely to be affected by marine or lacustrine processes only during large storms. Large-scale aerial photographs and topographic maps of these interior areas are usually available and are adequate for study purposes.

Information on the active, more variable shore and shoreface zones is usually obtained by direct field survey. The preferred surveying technique involves collecting a series of shore-normal profile lines. The profile lines should extend landward of the influence of inundation by moderate storms, usually behind frontal dunes. The preferred closure depth is at the toe of the shoreface, often defined as a selected depth contour where variability becomes minimal. Profile lines should be spaced at intervals close enough to show any significant changes in lateral continuity. Profile lines are connected with a surveyed shore-parallel baseline from which position and elevations of each profile origin can be determined.

Resurveying control profile lines at selected intervals of time can reveal seasonal patterns. In addition, special surveys can be made after significant storms and events to determine whether these events affect the local beach system. As previously noted, however, lead time and expense do not usually allow for an extended period of monitoring profiles. At a minimum, summer and winter profiles are recommended to identify seasonal variability.

A permanent or semi-permanent benchmark, or set of benchmarks, is required for reoccupying a profile site over successive months or years. On a rapidly transgressing coast, these benchmarks should be located near the landward end of the profile line in order to minimize storm damage, although locations which might experience dune burial should be avoided. On a regressive coastline, the benchmarks can be placed closer to the shoreline. In both cases, care should also be taken to reduce the visibility of benchmarks so that they will not be damaged by vandals.

Onshore portions of profiles are surveyed using standard land survey techniques and equipment. Extending profile lines offshore beyond wading depths requires a boat or amphibious vehicle. Positioning offshore can be accurately and rapidly established by one of several high-precision navigation systems now available. Fathometers can be used for continuous profiling of the area seaward of the breaker zone but the signals are disrupted by breaking waves in the zone. Further, boats suitable for offshore use cannot approach close enough to the shore to connect directly to the land profile. Amphibious vehicles are better suited to this task because they can traverse the sea-land boundary and establish the continuity of profile lines.

In waters relatively close to shore during favorable weather conditions, one survey platform that is often used is a sea sled, consisting of a long upright stadia rod mounted vertically on a base frame with sled-like runners (Clausner, Birkemeier, and Clark 1986) or a mast with a prism for use by total station survey techniques (Fredette et al. 1990). The sled is towed, winched, or otherwise propelled along the profile lines (self-propelled, remote-controlled sea sleds are currently being developed) while frequent depth and position data are determined using onshore instruments. Because neither the sea sled nor the onshore survey and positioning equipment are floating, elevations are not subject to wave or tide variations, thus providing a more accurate comparison between repeated surveys. At present, bottom samples must be obtained from a boat or amphibious vehicle working in conjunction with the sea sled. The technique is currently limited to use within 4 km of the coast and water depths of 12 m, less than the height of the sled mast.

Sedimentological and Stratigraphic Techniques

In addition to examining large morphological features, examination of small-scale morphology can provide much information on the variable nature of coastal forms and processes. A number of techniques exist that enable collection of data on present and past processes at a site, using information on the characteristics of sediments and surface stratigraphy.

Knowledge of sediment characteristics may be useful for predicting sediment movement during storms, the nature of seafloor features, and the geologic history of the area of investigation. Sediment transportation is influenced by properties such as size, shape, and composition, with grain size being most important. Differential transport of coarse and fine, angular and rounded, and light and heavy grains leads to grading. There are several samplers for determining the movement of sediments (McCave 1979; Seymour 1989). Sediment traps are a direct means of estimating sediment movement (Kraus and Dean 1987). Bed surface sediments are typically collected with grab samplers and then analyzed using standard laboratory procedures as described in other sources (Fredette et al. 1990; Buller and McManus 1979).

For acquiring undisturbed samples or samples at greater depths, some type of coring device must be employed.

Cores allow retrieval and examination of the subsurface material in the area of investigation. From the recovered sediment sequence, much information regarding history of the depositional environment and processes can be determined. Depending upon the information required, the types of analyses performed on the core may include grain size, sedimentary structures, the occurrence of shells and minerals, organic content, microfaunal identification, x-radiographs, age dating, and engineering tests. If it is important to understand much about the geologic history of a site, vibratory corers will be necessary, which may retrieve samples exceeding 12 m in length. Even deeper cores can be recovered with rotary drilling equipment. If only information regarding recent processes is necessary, then a box corer, which samples up to 0.6-m depths in the sediment and provides detailed information regarding sedimentary structures, will likely be adeqate.

Standard surveying techniques or large-scale aerial photographs are preferable to side-scan sonar for acquiring bedform data on exposed sand banks at low water. Dimensionless parameters of ripples and other bedforms can indicate depositional environment (Tanner 1967). Flow directions can be assessed in terms of the trace of the crestline (Allen 1968). Wave-formed structures reflect the velocity and direction of the oscillatory currents, plus the length of the horizontal component of orbital motion and the presence of velocity asymmetry within the flow (Clifton and Dingler 1984). The flow strength for inter-tidal estuarine bedforms can also be estimated for a given flow depth by a velocity-depth sequence of bedform development (Boothroyd 1985).

Heavy minerals may provide information regarding sources, processes, and other aspects of geomorphic variability in the coastal zone. Pronounced seasonal variations in heavy minerals may also occur in the beach and nearshore samples, with foreshore samples showing higher concentrations in winter than summer, and samples outside the surf zone showing lower concentrations in winter than summer. An explanation for this phenomenon is that light minerals are transported from the beach foreshore to deeper water during the winter, and are transported back again during the summer (Inman 1953; Nordstrom and Inman 1975).

Physical Models

The use of physical models can also shed light on the geomorphic variability of coasts. Physical models require scaling and calibration, and significant time and expense to set up initially. Once in operation, however, they allow for direct measurement of process elements, and the study and isolation of variables that are difficult to assess in the field. Some examples of physical

model experiments, conducted principally in wave tanks, that help elucidate geomorphologic variability of coasts include studies of littoral drift blockage by jetties (Sireyjol 1965), breaker type classification (Galvin 1968), experiments of cliff erosion (Sunamura 1983), relationships of storm surge or short-term water level changes to beach and dune erosion, and studies of suspended sediment concentration under waves (Hughes 1988). Physical models are considered invaluable for many coastal engineering studies. (A detailed description of the types and results of such models is beyond the scope of this report.)

Numerical Models

The use of numerical models in assessing changes in coastal geomorphology is rapidly increasing in sophistication. Models include those that perform wave refraction and longshore transport computations, those that estimate beach profile response and coastal flooding, and those that examine shoreline change and storm-induced beach erosion (Dean and Maurmeyer 1983; Komar 1983; Birkemeier et al. 1987; Kraus 1990). While a detailed description of such models is beyond the scope of this report, such studies can greatly assist the understanding of coastal processes and landforms in the vicinity of a study site. In turn, prior characterization of local geomorphology based on independent data sources can provide an invaluable check on the reasonableness of such models' results.

6 Summary and Conclusions

Coastal environments show great geomorphic diversity over space and over time. Spatial diversity occurs because coastal landforms develop in a variety of terrestrial and marine environments, composed of materials that include a variety of rocks and sediments. In addition, environmental factors, such as coastal winds, waves, tides, currents, storms, sea level, tectonics, and sediment supply show geographical variation. Temporal diversity in landforms and materials occurs largely because environmental factors show temporal variations. Temporal variations may be cyclic, noncyclic, or unidirectional over the time period examined. The geomorphic variability of coasts reflects the multiplicity of geomorphic and geologic responses over a variety of time scales.

Types of geomorphic zones include (a) beaches and the nearshore zone, (b) coastal dunes, (c) the shoreface, (d) inlets, (e) shelf shoals, (f) deltas, (g) estuaries, (h) reefs, (i) mudflats and mangroves, (j) strand plains, and (k) cliff coasts. It is an important aspect of engineering and geologic studies to assimilate and interpret evidence regarding geomorphic variability over the multiple time scales that occur in these wide-ranging coastal environments.

Study of the geomorphic variations of coasts can be approached over a variety of time scales. Three principal time scales that are important in assessing geologic and geomorphic changes in coasts include the following: (a) modern studies based largely on field data or laboratory and office experiments regarding environmental processes; (b) historic studies based largely on information from maps, photography, archives, and other sources; and, (c) paleoenvironmental studies based largely on stratigraphy and associated geological and paleoenvironmental principles. In actuality, however, these general time scale approaches show overlap. Further, within each of the categories, certain time scales may be of particular importance for influencing coastal changes. For example, tidal and seasonal changes are significant in modern studies and Holocene sea level history is important in paleoenvironmental studies.

Many coastal geomorphic features are temporally variable and tend to change form with changes in certain critical environmental factors. Some of these changes are cyclic and relate mainly to seasonal variations in wave climate; others are the result of rare intermittent events such as major storms

or long-term unidirectional trends that may be climatic or related to changes in relative sea level and/or sediment supply. The planning and design of coastal engineering projects and the long-term management of coastal areas require a basic knowledge of the likely geomorphic variations that can be expected to occur during the lifetime of the project. Thus, before planning and designing, some study of the environmental and geomorphic features of the coastline to be engineered, and its relation to adjacent coastlines, should be given detailed attention.

Many types of processes occur in the coastal zone. Inland of the coastline, terrestrial processes are dominant, except during severe storm surges and where the coastline is on a barrier backed by marginal marine features such as lagoons, sounds, bays, and marshland. The shore and shoreface are the main focuses of coastal marine processes: waves, currents, tides, and storm surges. Seaward of the shoreface, inner Continental Shelf features such as linear and arcuate shoals may change because of interaction with large waves and storm-generated shelf currents. Other important causes of change in coastal zone features are changes in relative sea level, increase or decrease of sediment supply, and construction of engineering projects. Processes acting on consolidated rock are usually much slower to bring about significant morphological change, and some features may remain relatively unchanged for centuries. Also, some changes such as relative sea level may take place so slowly that they are difficult to detect and measure.

Data collection on temporal changes in coastal features and environmental factors is often difficult and expensive. Ideally, time-series data should cover periods of several years, but this is rarely possible because of required lead time and funding constraints. In some cases, there are historical resources in the form of earlier hydrographic surveys, aerial photography, and prior studies that can provide information about past temporal changes, but many are deficient in detail, quality, time span, or frequency of observations and measurements.

It is important to obtain data on the processes, such as waves and currents, that affect coastal features and, insofar as possible, gain an understanding of the connection between processes and variable geomorphic features. Many instruments are now available for acquiring time-series data on waves, currents, winds, and other dynamic factors of the environment. Although capable of obtaining high quality data, these techniques of data collection can be expensive if used alone. With good historical sources such as weather charts, much can be learned concerning factors such as waves and storm surges by hindcasting, and instruments may be more economically used in a limited role to verify data. The difficulty in measuring many environmental factors is primarily economic rather than technical.

The geomorphic variability of the coast requires that a range of factors be considered in the management and engineering of coasts for storm protection and navigation. Environmental and geomorphic factors cause the rates, scale, and nature of change in coastal environments to be highly variable, and the

interaction of these various factors is generally complex. Coastal scientists and engineers will be more successful in planning and designing coastal projects by determining the geomorphic variability of the coastal zone over short and long time scales.

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Appendix A
Glossary of Geologic Terms

ANGLE OF REPOSE The maximum slope at which cohesionless sediments are stable.

ARCH A rock feature in the form of an arch lying close offshore. The form and isolation are products of marine erosion.

AVULSION A tearing away. Often refers to a rapid change in a river's course.

BACK BARRIER Pertaining to the lagoon-marsh-tidal creek complex in the lee of a coastal barrier island, barrier spit, or baymouth barrier.

BARCHAN DUNE A dune of crescentic shape with the convex side facing the prevailing wind.

BEACH CUSPS One of a series of low mounds separated by crescentic-shape troughs spaced at regular intervals along a beach.

BEDFORMS Any deviation from a flat bed that is readily detected by eye, e.g., ripple marks, sand waves.

BIOEROSION Erosion of rocks or sediment created by biological activity, especially boring organisms.

BIOGENIC Of biological origin. Usually sediments composed of the hard parts of plants or animals and organic reef masses.

BIOTURBATION The disturbance of sediment bedding by the activities of burrowing organisms.

BLOWOUTS Places in vegetated dunes or sand flats where bare spots occur and are subjected to wind erosion.

BRYOZOA Invertebrate belonging to the Phylum Bryozoa and characterized by colonial growth and a branching, twiglike skeleton.

BYPASSING The movement of sediment across a natural or manmade barrier to alongshore sediment movement.

CARBONATION A chemical weathering process involving the transformation of calcium, magnesium, potassium, and sodium minerals into carbonate and bicarbonates of these metals by carbon dioxide contained in water.

CHELATION The taking up or release of a metallic ion by an organic molecule.

CLASTIC Sediment or rock composed of particles of pre-existing rocks or minerals that have been transported out of the place of origin.

CONTINENTAL SHELF The broad shallow submarine plain that fringes many continental coasts.

CORALLINE Pertaining to the large group of hard, calcareous, external skeletal, bottom-dwelling marine coelenterates of class Anthozoa.

CORIOLIS EFFECT The apparent deflection of moving objects from a straight path caused by earth rotation. Moving bodies appear deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere.

CREVASSE Splay deposit Deposits laid down by river water emerging from a break or rupture in a river's levees. The deposits sometimes resemble a small delta.

CYCLONIC Pertaining to an atmospheric system that is characterized by low pressure and counterclockwise winds north of the equator and clockwise winds to the south.

DIAGENESIS Changes undergone by sediments after their initial deposition. The term often refers to compaction, cementation, and replacement.

DIAPIR Structure in which the core of an anticline breaks through overlying rocks.

DIASTROPHISM The deformation of large masses of rock.

DIP The angle of a bedding plane to the horizontal.

DISTRIBUTARY A branch of a stream flowing out of another stream. Distributaries are most common in deltas.

DOLOMITIC Containing dolomite, a calcium magnesium carbonate. Dolomite is both a mineral and rock name.

DRUMLIN A glacial feature usually consisting of a large mound of glacial drift.

EBB TIDAL DELTA Shoal formed on the seaward side of inlets by ebb tidal currents.

ECHINOIDS A class of free-moving echinoderms, mostly with rigidly plated bodies.

EUSTATIC Refers to worldwide changes in sea level due to changes of the volume of water in the ocean or a change in the volume capacity of the basins.

FJORD A glaciated valley that has been partly drowned by relative sea level changes or as a result of overdeepening by the glacier that formed it.

FLOOD TIDAL DELTA A shoal complex on the landward side of an inlet deposited by flood tidal currents.

FORAMINIFERA Protozoans characterized by tests of one to many chambers composed of calcite or of agglutinated particles.

FRIABLE Weakly consolidated or weathered rock that is easily broken up.

GEOMORPHIC Pertaining to landforms.

GLACIAL DRIFT Sediments associated with glacial deposition.

GLACIOISOSTATIC Vertical crustal movement including depression of a land area due to the weight of overlying glacial ice or elevation of the land due to unloading of glacial ice.

GLACIOLACUSTRINE Pertaining to glacial deposition or erosion in a lake.

GROINS Low structures oriented perpendicular to the shore. They are intended to trap or retard alongshore movement of beach material.

HOLOCENE EPOCH The latest epoch, which began about 10,000 years before present, following the last glaciation.

HYDRATION Absorption of or chemical reaction with water.

HYDROLYSIS A decomposition reaction involving water, usually between silicate minerals and aqueous solution.

IGNEOUS Said of rock or minerals which solidified from molten or partly molten materials. Granite is a common example.

ISOBARS Lines of equal barometric pressure on a weather map.

ISOHALINES Lines of equal salinity on a chart of saltwater bodies.

KARST A limestone terrain of very irregular topography due to extensive solution of the limestone rock.

KATABATIC WIND Any wind blowing down an incline. If the wind is cold, it is known as a foehn; if it is cold, it may be a fall or gravity wind.

LACUSTRINE Pertaining to a lake.

LAGOON A shallow body of water behind a coastal barrier island, barrier spit, or baymouth barrier. Also applied to the water bodies inland of coral reefs and atolls.

LAPSE RATE The adiabatic rate of change of a meteorological element (such as temperature) associated with a change in height.

LITHIFIED Refers to rocky or hardened sediments.

LITHOLOGIC Pertaining to the character of rock.

LITHOLOGY The general characteristics of a rock or sediment.

LITHOSPHERE The solid portion of the earth, usually referring to the crust and upper part of the mantle (about 100 km in thickness total).

LONGSHORE CURRENT A current flowing alongshore, primarily inshore of the outer breaker line.

LONGSHORE DRIFT Sediment moving alongshore on the foreshore and adjacent breaker zone due to waves and wave-generated currents.

MARGINAL SEAS A partly enclosed sea bordering a continental landmass.

MASS WASTING The movement of soil and sediments downslope due to gravity.

MORAINE Accumulation of glacial till left by melting of glaciers. Material is deposited directly by the ice.

MORPHODYNAMIC A concept introduced by Wright, Short, and coworkers which represents the integration of seemingly disparate hydrodynamic and morphologic factors into a coherent morphologic model with distinct states or stages.

MORPHOLOGY In geology, the visual shape of landscape features, either singly or as a group in a given area.

NATURAL LEVEE A natural embankment along the shore of a stream deposited by overflow during floods.

ORBITAL WAVE MOTION The orbital displacement of water as a wave passes.

ORTHOGONAL A line perpendicular to a wave crest.

OOLITE Calcareous rounded particles ranging from egg-shaped to button-shaped that are precipitated directly from sea water.

PALEOENVIRONMENT An ancient environment that is reconstructed by the evidence of fossils, bedding and sediment, or rock characteristics.

PALEONTOLOGY The study of past geological history, primarily by analysis of fossil remains of organisms.

PALUDAL Pertaining to low water-covered land, such as swamps and marshes.

PALYNOLOGY The study of spores and pollen. Often used to help reconstruct past environments.

PARALLAX The apparent displacement of an object due to a change in the position of the observer.

PEDOLOGY The science of soils.

PHOSPHATIC Containing phosphate, usually in combination with other elements, to form minerals.

PLEISTOCENE A subdivision of the Quarternary period. The Pleistocene encompasses the ice ages.

QUATERNARY The period following the tertiary period and containing the Pleistocene and Holocene epochs.

RADIOCARBON DATE Age of a carbonaceous material derived from the radioactive changes in carbon 14 through time.

RADIOMETRIC Chronological age determined by study of suitable radioactive substances.

RELATIVE SEA LEVEL The level of land and sea in respect to one another. Changes in relative level may be due to eustatic sea level changes or the vertical movement of the landmass.

SALTATION The movement of sediment particles in a series of short jumps from the bottom.

SALT DOME A diapir of salt that breaks through overlying rocks or sediment. Hydrocarbon-bearing strata are often associated with salt domes.

SEDIMENTOLOGICAL Pertaining to the science of sedimentary rocks and their formative processes.

SEISMIC REFLECTION A geophysical method of obtaining sonic representations of subbottom stratification. Seismic reflection surveys are performed on land and at sea.

SELECTIVE SORTING The process in sediment transport of differential response of sediment particles due to differences in size, shape, and specific gravity.

SETUP Elevation of a water surface due to onshore mass transport of water by wind and waves.

SHORE The border of a landmass with the sea.

SHOREFACE A sloping bottom zone between the shore and the inner continental shelf.

SILICEOUS Containing silica.

SINK A process that subtracts sand from a littoral compartment.

SLIP FACE The steeply sloping lee side of a dune or sand wave.

SPIT A long narrow ridge of sand usually extending from a headland parallel to the general trend of the coast.

STACK A rock column isolated offshore by coastal erosion.

STRATIGRAPHIC Pertaining to the study of stratified rocks and sediments.

STRIKE The orientation of the line of intersection of a bedding fault plane with a horizontal plain.

STORM SURGE A rise of water level above normal due to wind stress on the water surface.

SUBAERIAL Pertaining to the land surface as contrasted with subaqueous, i.e., below the water surface.

SUPERPOSITION A geological concept that a rock or sediment overlying another layer is younger in origin, provided no structural deformation exists that would reverse the sequence.

TECTONIC Pertaining to structural movements of rock masses.

TECTONIC PLATES A large part of the earth's crust that moves as a unit with respect to other plates.

TEPHRA Clastic material that has been ejected from a volcano.

TERRESTRIAL Pertaining to land as opposed to the ocean.

TILT Sideways inclination of an aircraft or spaceship.

TSUNAMI A long-period wave caused by an underwater disturbance such as an earthquake. Tsunamis are often highly destructive if they reach a coast.

TIDAL PRISM The total amount of water that flows into a lagoon, harbor, or estuary and out again with the rise and fall of the tide.

WAVE STEEPNESS The ratio of wave height to wave length.

WEATHERED The condition of rocks that have undergone physical and/or chemical alterations due to exposure to the elements.

YAW Refers to an aircraft's or spaceship's turning by angular motion about a vertical axis.

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